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THE ORIGIN AND NATURE OF BRECCIATION OF DEVONIAN STRATA IN
WOOD BUFFALO NATIONAL PARK

by



Darrell G. PARK

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH
IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE
OF MASTER OF SCIENCE

Geology

EDMONTON, ALBERTA

FALL, 1984

THE UNIVERSITY OF ALBERTA
FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled THE ORIGIN AND NATURE OF BRECCIATION OF DEVONIAN STRATA IN WOOD BUFFALO NATIONAL PARK submitted by Darrell G. Park in partial fulfilment of the requirements for the degree of Master of Science.

ABSTRACT

The Devonian Elk Point and Beaverhill Lake groups in Wood Buffalo National Park were deposited on the extreme northern edge of the La Crete Basin. Interruption of shallow water sedimentation produced numerous depositional breaks.

At Salt River, the Lower Keg River Member occurs beneath the Keg River marker. This member is subdivided into units A to C, the sediments of which were deposited in a shallow subtidal to supratidal, platform environment. A diastem exists between units B and C. During this time interval, subaerial exposure produced extensive diagenesis.

Two brecciation events affected the Lower Keg River Member. The first phase of brecciation involved pseudoanticlines that formed in response to subaerial exposure of units A and B. Submarine pseudoanticlines and tepees formed in units A, B and C. Modern pseudoanticlines form along coastlines. The pseudoanticlines and tepees in the Salt River strata indicate a coastal depositional setting. The second brecciation phase produced laterally restricted collapse breccias. The previously formed tepees acted as sites for further dissolution, resulting in collapse of the overlying Upper Keg River Member.

Evaporites of the Fort Vermilion Formation, limestones and dolostones of the Slave Point Formation and shales of the Waterways Formation (Peace Point Member) outcrop at Peace Point. An angular unconformity separates the Fort Vermilion Formation from the Slave Point Formation.

The Slave Point Formation comprises sediments deposited in a very shallow water to supratidal environment. During subaerial exposure of the Slave Point Formation, prior to deposition of the Peace Point Member, a karst profile (capped by a calcrete) developed in the underlying strata. Extensive diagenesis associated with the karst/calcrete profile produced the different lithofacies and breccias along Peace River. Further diagenesis beneath a pre-Cretaceous unconformity resulted in later breccias.

At least three phases of brecciation have affected the strata around Peace Point. During collapse 1, solution breccias were produced prior to deposition of the Peace Point Member. Collapse 2, the major brecciation event, produced evaporite-solution-collapse breccias after deposition of the Peace Point Member and prior to Cretaceous sandstone deposition. Collapse 3 represents a post-Cretaceous but pre-glaciation event that produced a sandstone breccia.

Cretaceous sandstones, siltstones and shales in the sandstone breccia contain faunal evidence of the Devonian-Cretaceous unconformity as well as lithologies and faunas typical of the McMurray Formation. The occurrence of these clastic rocks indicates that Wood Buffalo National Park was once covered with Cretaceous rocks that were later eroded.

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I. INTRODUCTION

This thesis concentrates on brecciated strata which occur in the Keg River Formation, the Fort Vermilion Formation and the Slave Point Formation of Wood Buffalo National Park. The breccias commonly contain rocks which have obviously been derived from younger strata. Such rocks may or may not be representative of strata presently outcropping in Wood Buffalo National Park. Therefore, in order to understand the origin of the breccias it is necessary to be cognizant of the Devonian and Cretaceous stratigraphy of northern Alberta, as well as in Wood Buffalo National Park.

A. LOCATION

The breccias and other rock types described in this thesis occur along the Salt River, to the south of the Salt River Bridge (Fig. 1) and in sequences that are exposed along the Peace River in a 4 km stretch west of Peace Point and a 1.5 km stretch east of Peace Point (Fig. 2).

B. OBJECTIVES

The purpose of this study was to examine the breccias, their geological setting and to ascertain the geological history of the rocks exposed along Salt River and Peace River. The main objectives of the study were: 1) to determine the nature and genesis of brecciation, 2) to determine the age of the brecciation, 3) to identify

unconformities, their nature, and how they relate to the brecciation processes and 4) to provide a better understanding of the correlation of the strata in Wood Buffalo National Park with the rest of northern Alberta.

C. FIELD METHODS

Work along the Salt River and Peace River was conducted in the summer of 1982 (August 13 - September 3) and in the summer of 1983 (July 26 - August 3). Only one day out of the combined field seasons was lost to rain. A complete set of field notes related to this thesis is housed with the Paleontological Collections, Department of Geology at the University of Alberta.

Field methods involved detailed description of the exposed strata. Individual units were described and approximately 500 lithologic and fossil samples were collected. All lithological samples were cut and slabbed and approximately 200 thin sections were made. The study was aided by 300 photographs (both black and white, and color) that were taken.

Most of the localities were reached by walking and driving, however a canoe was periodically used for transportation along the Peace River. Helicopter support was supplied on one occasion by the Fire Control personnel of Wood Buffalo National Park.

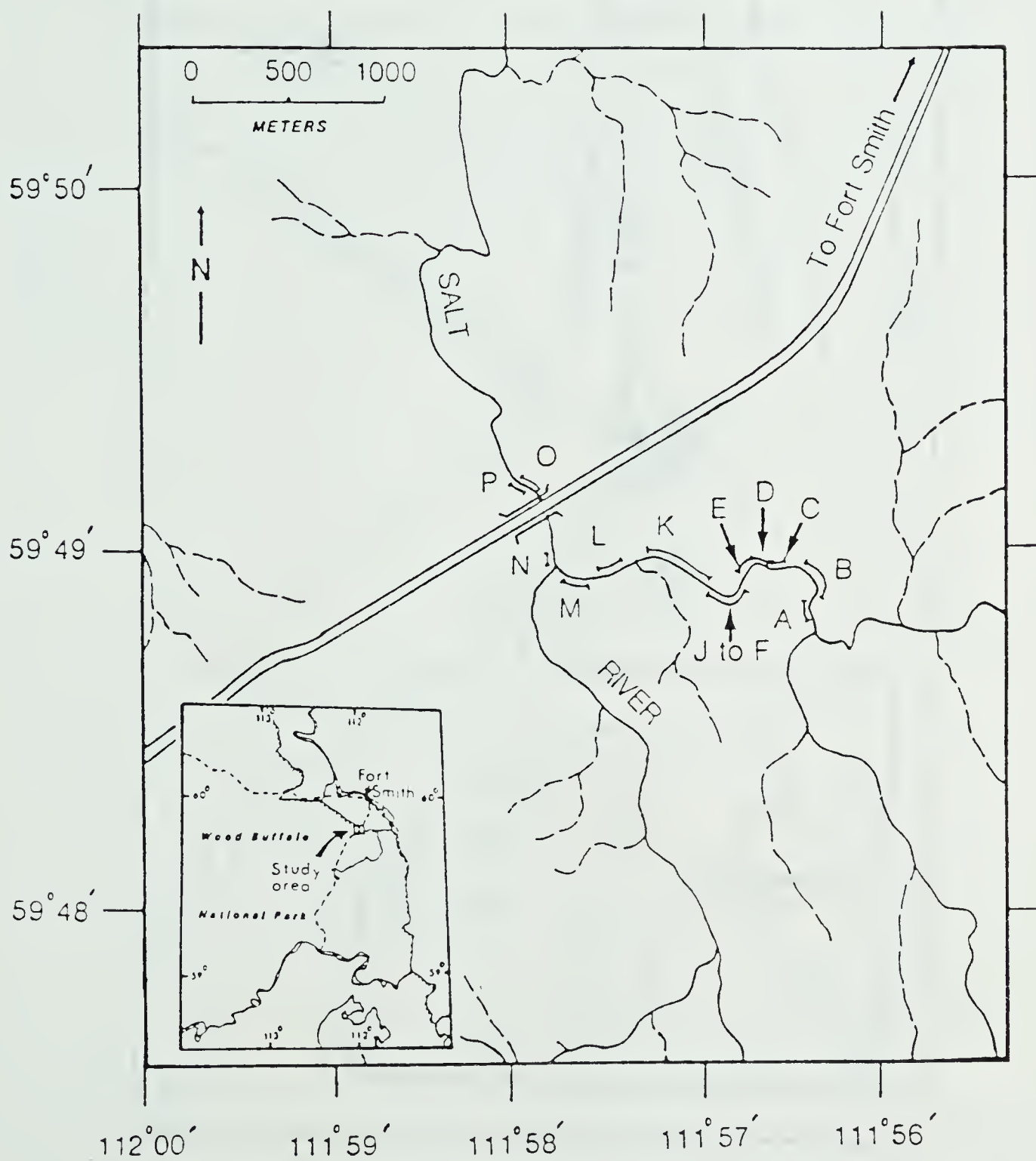


FIGURE 1. Sketch map showing the location of sections measured along Salt River.

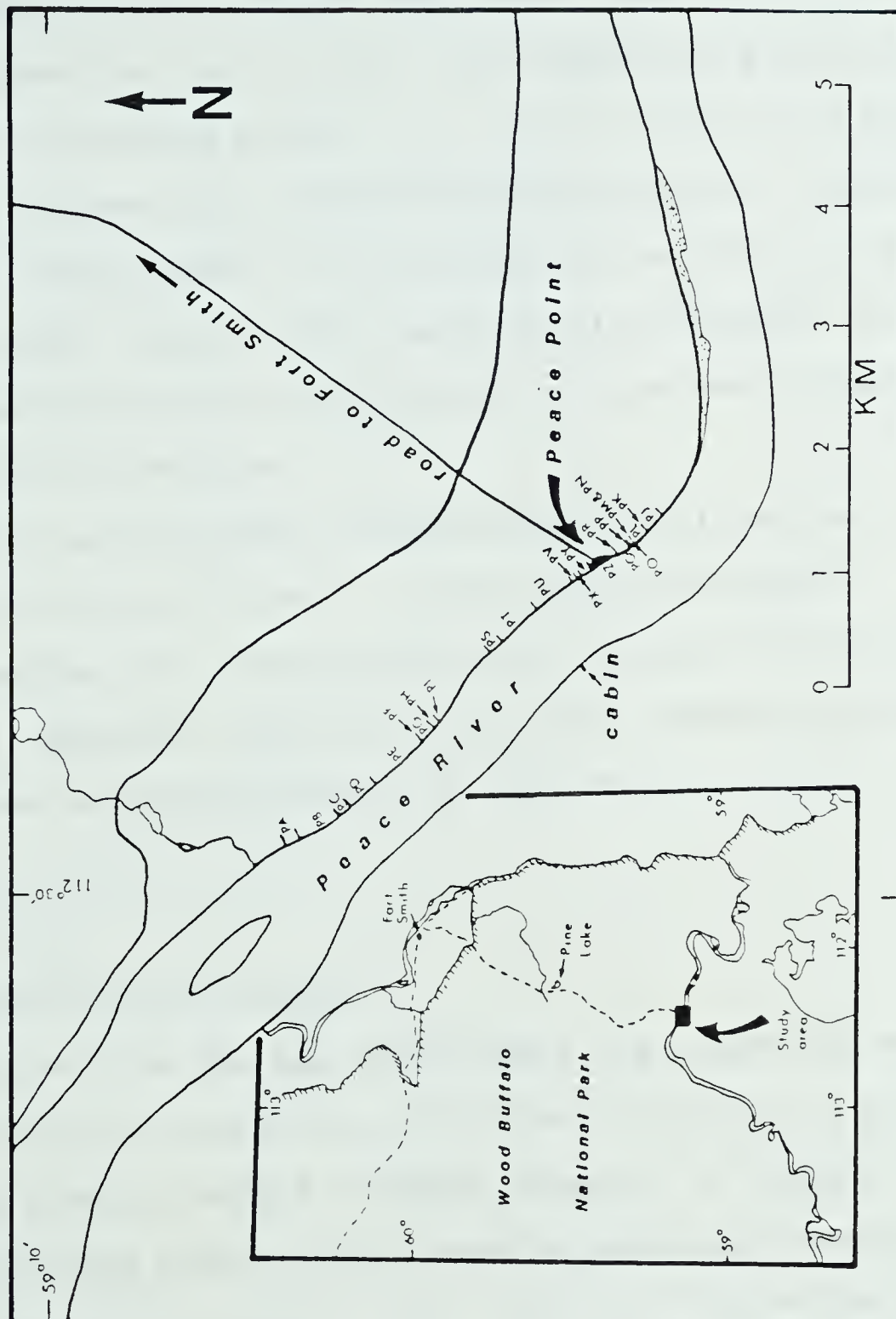


FIGURE 2. Sketch map showing the location of sections measured along Peace River.

D. PREPARATION OF ROCK SAMPLES

Preparations of the rock samples collected were needed for subsequent description and classification.

Slabbed portions of the rock samples were polished with a 400 grit grinding powder on a circular lath. The surface of the slabs was not coated with plastic or any laquers. Glycerol, when added to the surface of the slab, provided a smooth, clear coating that could easily be washed off. For photographic purposes the glycerol was an easily applied, non-reflective medium.

Thin sections were treated with a multipurpose stain, a mixture of Alizarin red - S stain and a Potassium Ferricyanide stain which permits the identification of calcite, dolomite, ferroan calcite and ferroan dolomite. The stain was mixed and applied as described by Dickson (1965, 1966).

E. CLASSIFICATION SCHEMES

Rocks from the two study areas are comprised mainly of carbonates and evaporites. The classification of these rocks is relatively straight forward. However, at Peace Point, carbonate and clastic constituents are mixed thereby creating a more complex rock. Like any classification scheme there is always the problem of 'pigeon-holing' the rocks and subsequent boundary selection between adjacent fields. The rocks were initially classified using a non-restricted triangular classification scheme (Fig. 3; modified from

Jones, 1974) where the three components are quartz, calcite and dolomite. Ten categories of rock type were established by subdivision of the triangular diagram (Fig. 3) This triangular classification scheme allowed the initial classification and separation of the rocks whereas other classification schemes (Folk, 1959, 1968) were used to further separate the lithologies.

SANDSTONES

Using the triangular diagram (Fig. 3) clastic rocks were separated from carbonate rocks. The sandstones were further separated based on their calcite content. Some of the sandstones contain up to 30% calcite. Those sandstones with $> 20\%$ calcite are termed calcareous sandstone, whereas, sandstones have $< 20\%$ calcite.

Clastic rocks plotting in the shaded portion of triangle A (Fig. 3) were further classified according to Folk's (1968) sandstone classification. The terms "mature" and "immature" were used to classify the clastic rocks, based on textural maturity. "Mature" and "immature" refer to those sandstones that contain $< 15\%$ and $> 15\%$ clay matrix, respectively. The term "immature" does not carry the additional implication that the sand grains are poorly sorted and/or angular. The modal grain size, based on the Udden - Wentworth size grade scale (Wentworth, 1922), is used as a further modifier. Roundness and sphericity are discussed using the terms outlined by Powers (1953). Grain fabric delineation and charts for visual estimation of

sorting were those derived by Pettijohn et al. (1973).

CARBONATE ROCKS

Carbonate rocks were initially defined by using triangle A (Fig. 3). Rocks classified as limestone or dolostone consist of > 90% calcite or dolomite, respectively. If the percentage of the least constituent mineral (whether calcite or dolomite) was in the 49 to 89% range the rock was either calcareous or dolomitic, respectively. Following determination of the dolomite/calcite percentages the rocks were classified according to crystal size of the major carbonate constituent, based on Folks' (1959, p. 16) crystal/grain size scale for carbonate rocks (Fig. 3). For example, in a dolomitic limestone the modal calcite crystal size would determine the modifier to be used for the rock name. In this study, the grain size is the same for both the dolostones and the limestones. Unlike Folks' (1959, p. 16) classification, distinction between those grains transported and those that are authigenic was not made.

Carbonate classification, in terms of fossil abundance, is important in the Salt River area, but negligible in the Peace Point area. No modifier was used if the fossils comprise less than 10% of the rock. However, rocks with 11 to 49% fossils were named "fossiliferous", while rocks with > 50% fossils were termed "highly fossiliferous".

Other constituents such as oncolites, pellets, or textures such as "rubbly", were used as modifiers as

necessary. To be used as a modifier the constituent must comprise > 15% of the rock.

Classification schemes such as Folks' (1959, p. 14) classification of carbonate rocks and his carbonate textural spectrum, (Folk, 1962, p. 76) were used where applicable. For example, the lithology of the marker unit in unit C (of the Keg River Formation) is best termed a packed biomicrite. All the lithological and textural inferences suggested by this rock name are as described by the original author.

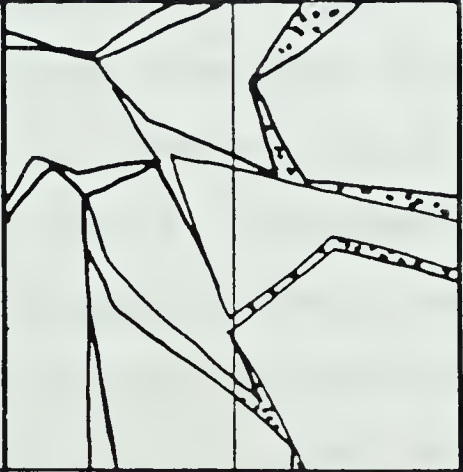
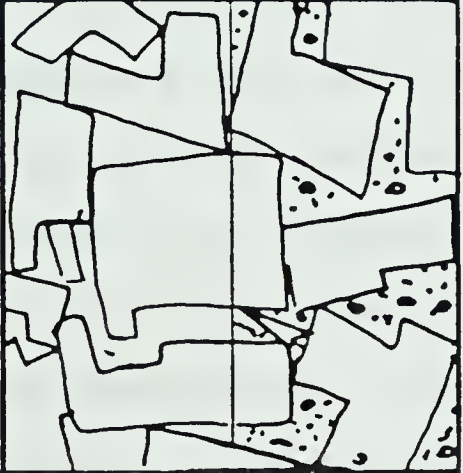
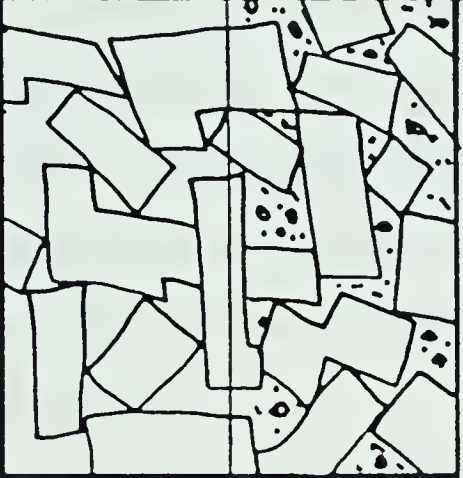
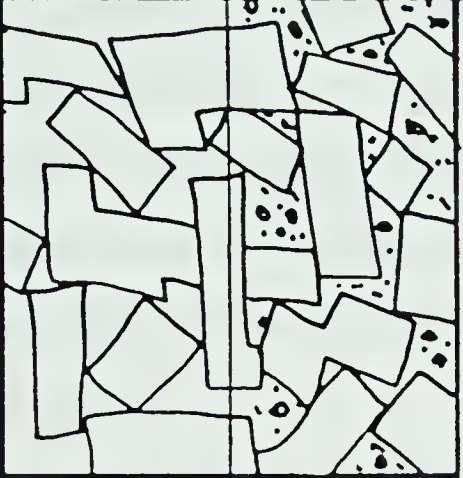
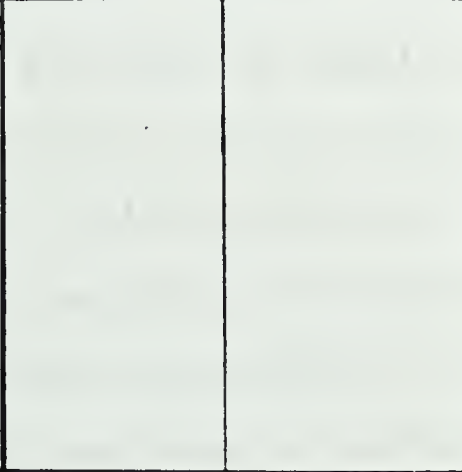
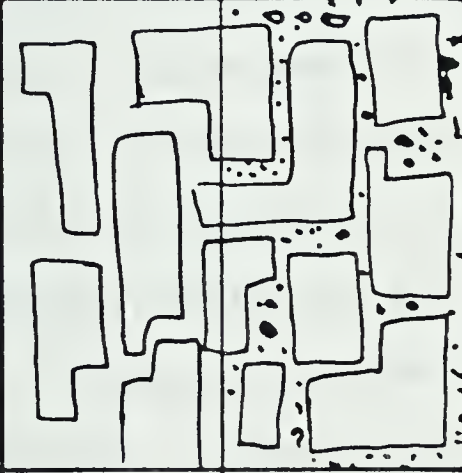
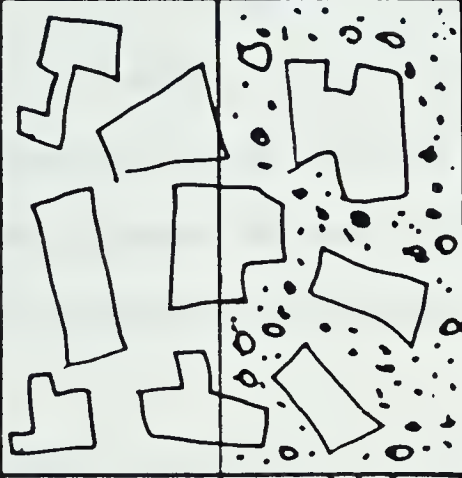
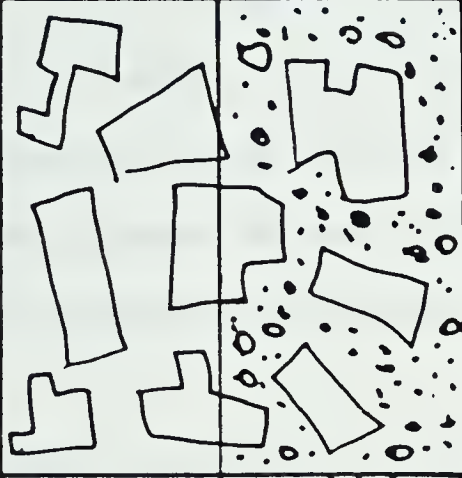
Extensive alteration of the carbonates by later diagenetic events has completely obliterated any original lithologies or textures of the rock. If oncolites, intraclasts, or in many cases, fossils did originally comprise a good portion of the rock they are now generally not recognizable. Carbonate classifications by Folk are such that the nature of the calcite in a carbonate rock should be known (i.e. a sparry calcite cement), with the genetic inference of the calcite being applied. Based on carbonate rocks from this study it is difficult to discern a sparry calcite cement from a sparry calcite that is the result of extensive aggradation. Therefore, the chosen classification scheme simply states the crystal size, devoid of genetic origins of the calcite. Also, inferences of maturity based on textures, as is done with Folk's (1962) classification scheme cannot be applied to a major proportion of the rocks.

BRECCIAS

The breccias consist of clasts (> 1 cm) and matrices that are formed of carbonates, evaporites, shales or sandstones. Lithological descriptions of the breccias are given, but a classification scheme based solely on lithological parameters has not been derived. Instead, if the breccia contains a prominent rock type or has a characteristic lithology, different from the other breccias, it was applied as a modifier to the breccia. For example, the "gypsum breccia" (lithofacies 6) is formed mainly of "gypsum" blocks. The "shaly breccia" has a relatively high "shale" content compared to the other breccias in the area.

Classification of breccia fragments based on spatial relationships and fragments was proposed by Morrow (1982). In this classification, the interfragment space, the mutual orientation of the fragments and the mutual proximity of the fragments are all considered (Fig. 4). Morrow's classification allows a purely objective, descriptive analysis of the breccias and has thus been adopted in this study.

Although Morrows' (1982) classification is the best nongenetic, descriptive breccia classification to date, it has one major pitfall. The reason it was not the only classification used is that the fabric spectrum, crackle-mosaic-rubble, represents the relative degree of disturbance observed as bedded strata becomes brecciated. Virtually all of the combinations of breccia rock names

Packbreccia		
<i>crackle</i>	<i>mosaic</i>	<i>rubble</i>
<i>open or cemented</i>		
<i>particulate</i>		
Floatbreccia		
<i>not applicable</i>	<i>mosaic</i>	<i>rubble</i>
<i>open or cemented</i>		
<i>particulate</i>		

Morrow (1982)

FIGURE 4. Breccia Classification Scheme.

produced by this classification can be found in a few meters of outcrop. This is certainly the case on the edges of collapse where nondeformed strata become progressively more brecciated laterally. The rock can change from a particulate crackle pack breccia to a particulate rubble float breccia. Morrow (1982) recognized this problem and also stated that many breccia fabrics indicate the existence of particular conditions that occur during the development of breccias of diverse origins.

Most breccias have been described in genetic terms with no descriptive classification applied; for example, a fault breccia, or solution collapse breccia, a forereef breccia, an intraformational breccia. However, determination of the mode of origin of a breccia is commonly difficult and subject to controversy; therefore, the use of a genetic classification is fraught with problems. In this thesis the breccias have also been described in terms of genetic origins, but the stratigraphic setting, lateral relationship to other strata, lithology, spatial relationships of the breccia fragments, relationship to unconformities and geologic history of the area have been considered first, both substantiating and justifying the genetic connotations placed on the breccia.

F. PREVIOUS WORK IN WOOD BUFFALO NATIONAL PARK

Middle and Upper Devonian strata outcrop throughout Wood Buffalo National Park and adjacent areas (banks of the Slave River). Previous work on the Devonian Stratigraphy was concentrated in seven areas in the Park and along the Slave River, namely; 1) the west bank of the Slave River, where the La Loche Formation, Chinchaga Formation and the Keg River Formation occur in a discontinuously exposed sequence (Camsell, 1917; Cameron, 1918, 1922; Norris, 1963; Craig et al., 1967; and Rice, 1967); 2) the Salt River where the Keg River Formation and the Chinchaga Formation occur (Camsell, 1917 and Tsui, 1982); 3) west of Salt River, along a 30 m high escarpment that overlooks the Salt Flats, where the Chinchaga and the Keg River Formations outcrop (Cruden et al., 1982; Tsui, 1982). The same formations were also studied at the Parson Lake Fire Tower by Cruden et al. (1982) and Tsui (1982); 4) Little Buffalo River and just outside the northern park boundary, at Little Buffalo Falls, where the Chinchaga Formation is overlain by the Keg River Formation (Norris, 1963, 1965; Tsui, 1982); 5) the Walk-In Cave where Tsui (1982) described the Chinchaga and Keg River Formations; 6) The Gypsum Cliffs, on the banks of the Peace River, where the Fort Vermilion Formation, Slave Point Formation and Waterways Formation are exposed (Camsell, 1917; Cameron, 1922; Kindle, 1928; Cameron, 1930; Warren and Stelck, 1949; Crickmay, 1957; Halferdahl, 1960; Govett, 1961; Norris, 1963, 1965; Richmond, 1965; Norris and Uyeno,

1983) and 7) Lake Claire and the Birch River where the Slave Point Formation and the Waterways Formation were studied, respectively (Bayrock, 1972; Norris and Uyeno, 1981).

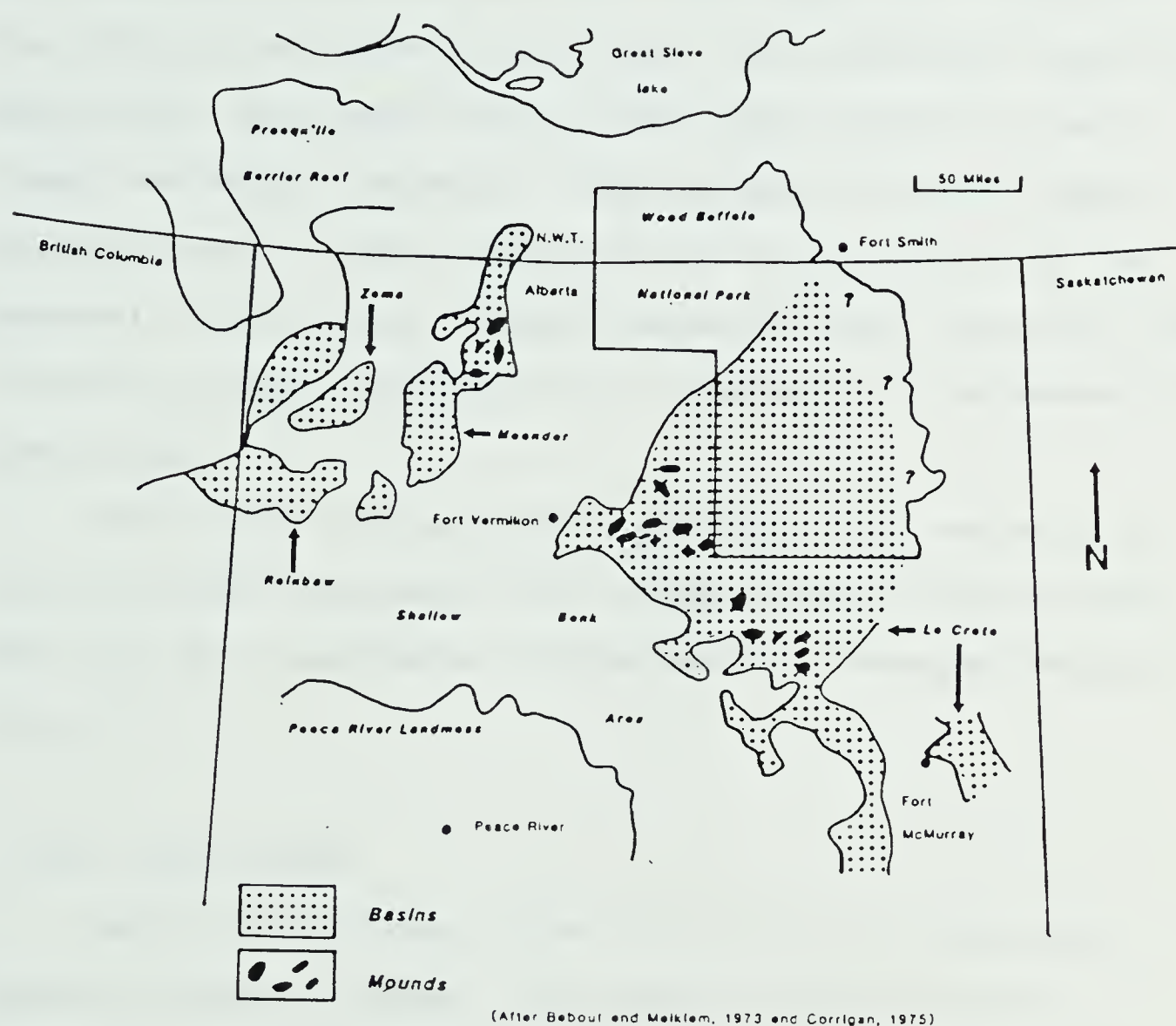


FIGURE 5. Middle to Upper Devonian Paleogeography of Northern Alberta.

II. GENERAL STRATIGRAPHY OF NORTHERN ALBERTA AND WOOD BUFFALO NATIONAL PARK

Apart from the Precambrian basement, isolated outliers of Cretaceous sandstone and surficial Quaternary deposits, the strata of Wood Buffalo National Park represent Devonian deposition. Such deposition in the Lower Devonian (Emsian Stage) and Middle Devonian (Eifelian and Givetian Stages) produced the Elk Point Group and the lower portion of the Beaverhill Lake Group. Deposition during Upper Devonian time (Frasnian Stage) resulted in the remainder of the Beaverhill Lake Group.

The strata in Wood Buffalo National Park and adjacent areas represent sediments that accumulated in the La Crete Basin and on a carbonate platform which surrounded the basin (Fig. 5).

A. ELK POINT GROUP

The Elk Point Group forms the base of the Devonian System in Western Canada. This group, represented by cyclical sequences of evaporites, carbonates and clastics, attains a maximum thickness of approximately 656 m (2,000 ft). Thicknesses of 492 m (1,500 ft) occur just south of the southern boundary of Wood Buffalo National Park; northwest of Lake Claire (approximate latitude 15° W longitude 59° N).

The Elk Point Formation was first defined by McGehee (1949) based on the evaporite/carbonate sequences exposed at Elk Point, Alberta. Later, Belyea (1952) raised the Elk

Point Formation to group status. Crickmay (1954) designated nine members (numbered 1 to 9) in the Elk Point Formation that occurs in the Anglo Canadian Elk Point No. 11 type well (2-21-57-5W4) as a more suitable "type section" of the Elk Point Formation. Law (1955), however, designated a new type well as the California Standard Steen River Well (2-22-117-4W6). He divided the Elk Point Group into four lithologic units, namely; the Chinchaga Formation, the Keg River Formation, the Muskeg Formation, and the Watt Mountain Formation. Subsequently, Van Hees (1956) split the Elk Point Group into upper and lower subgroups. He correlated his eight units with the nine members originally defined by Crickmay (1954). The eight units defined by Van Hees (1956) are (in ascending order), the Basal Red Beds, the Third Salt, Members 6 and 7, the Second Salt, the Ashern Formation, the Winnipegosis Formation, and the First Salt (Prairie Evaporite). Van Hees (1956) selected the base of the Ashern Formation as the boundary between the Lower and Upper Elk Point subdivisions. Later, Sherwin (1962) gave formation status to Van Hee's lower five units. He also redefined the base of the Upper Elk Point Sub-Group as the base of Crickmay's member 3. This was equivalent to Law's (1955) Keg River and Chinchaga Formations. Grayston et al. (1964) suggested that the division between the two subgroups should be drawn at the base of the first widespread correlatable horizon. They defined the base of the Winnipegosis/Keg River carbonate platform and the top of the

underlying Chinchaga evaporites as the contact between the Lower and Upper Elk Point Sub-Groups.

Corrigan (1975) stated that the boundary between the Upper and Lower Elk Point Sub-Groups is still not well defined for it is based on the lithological definitions of the units. In central and northern Alberta the Chinchaga Formation grades vertically into the Keg River Formation, making boundary delineation difficult. The boundary recognized by Grayston et al. (1964) and Corrigan (1975) is the boundary chosen in this study. It is placed at the level where the dark laminar carbonates, anhydrites and/or clastic mudstones of the Chinchaga Formation give way to the buff to brown dolomites of the overlying Keg River Platform.

Norris (1963) assigned the La Butte Formation, the Hay Camp Formation and the Unnamed Evaporites to the Elk Point Group. Craig et al. (1967) redefined Norris's (1963, 1965) formation names along the Slave River, allowing an easier application of regional terminology to the Slave River sections.

More recently the Elk Point Group terminology, as described by Grayston et al. (1964) and Corrigan (1975), has been applied to the Wood Buffalo National Park and Slave River Sections (Norris 1973; this thesis). The Lower Elk Point Sub-Group in the study area, comprises: the basal La Loche Formation, the Fitzgerald Formation, the Cold Lake Formation, and the Chinchaga Formation. The Upper Elk Point Sub-Group consists of the Keg River Formation and the

Nyarling Formation.

B. LOWER ELK POINT SUB-GROUP

LA LOCHE FORMATION

Unconformably overlying the Precambrian granitic gneiss and granites is an arkosic sandstone and conglomerate unit named the La Loche Formation (Norris, 1963). This formation outcrops at six localities along the Slave River (Norris, 1963,). At the type locality, along the Clearwater River, the formation is a pale brown, fine to medium grained, irregularly lenticular, thin bedded arkosic sandstone (Norris, 1973).

Along the Slave River, the basal beds of the formation are a depositional breccia, consisting of large angular fragments of granite in a sandy matrix (Norris, 1963). The unit fines upward into an interbedded sequence of very coarse grained arkosic sandstone, conglomeratic sandstone and sandy mudstone (Rice, 1967).

In Wood Buffalo National Park the lower contact of the formation is placed where the unaltered crystalline Precambrian basement changes to the detrital regolith of the La Loche Formation (Rice, 1967). The lower contact marks the initial transgression of the Lower/Middle Devonian Seas onto the Churchillian Precambrian highs.

Norris (1963) stated that the upper contact is gradational with the overlying carbonate Fitzgerald Formation. At the type section, the upper contact with the

McLean River Formation is placed where the lithology changes from sandstone to siltstone, argillaceous dolomite or evaporites. Sandy beds at the base of the McLean River Formation suggest that the contact is transitional in the type section area (Norris, 1973). A transitional upper contact with the Eifelian Fitzgerald Formation suggests that the La Loche Formation is either of Eifelian age or of an older Paleozoic age (Norris, 1963; Rice, 1967), however, the lack of fossils makes an age determination difficult. Along the Slave River, the La Loche Formation is diachronous (Craig et al., 1967) and is unconformably overlain by the Keg River Formation.

The La Loche Formation was assigned to the base of the Lower Elk Point Group solely because of its lithologic similarity with other basal redbeds that occur elsewhere in the Elk Point basins of Alberta and Saskatchewan. These redbeds have generally been assigned a Lower or Middle Devonian age. Norris (1973) stated that the basal redbeds rise in the section and appear to be younger in age moving from the west (deeper portions of the Elk Point Basin) towards the eastern margin of the Elk Point Basin (La Crete Basin). Grayston et al. (1964) tentatively assigned a Middle Devonian age to the Basal Redbeds of central and northern Alberta. They described them as an 82 m thick (269 ft maximum) unit consisting of red dolomitic shales, siltstones and sandstones that overlies the crystalline Precambrian basement. Crickmay (1954), Law (1955) and Sherwin (1962)

described basal redbeds resting unconformably on basement rocks. They also suggested a Middle Devonian age for the redbeds. Hamilton (1971) described redbeds overlying the Precambrian basement and underlying the Lotsburg Formation of the Lower Elk Point Group in central-northeastern Alberta. Norris (1973) also suggested that this basal redbed was equivalent to the La Loche Formation of northeastern Alberta. Both Hamilton (1971) and Norris (1973) suggested a Lower Devonian Age (Emsian) for the basal redbeds of the Clearwater-Athabasca area due to their stratigraphic position beneath the Emsian Lotsburg Formation. Consequently, they also assigned the La Loche Formation an Emsian age.

The Slave River of northeastern Alberta marks the very edge of the La Crete Basin. As deposition of the La Loche Formation approached the basin edge, along the western edge of the Precambrian Shield, it would have risen substantially in the stratigraphic section as it overlapped the Precambrian highs. Craig et al. (1967) suggested that the La Loche Formation was still being deposited during Givetian time as the transgressional seas pushed deposition further eastward on to the shield.

The La Loche Formation, is herein considered to be a diachronous unit spanning a minimal depositional time interval from Emsian to Givetian time (Lower to Middle Devonian). The La Loche Formation may be considerably older, but until the La Loche can be properly dated along the Slave

River the above ages will be used in this study.

FITZGERALD FORMATION

The type section of the Fitzgerald Formation is located between Caribou and Stoney Island on the Slave River. Norris (1963) measured a maximum of 8.9 m (27 ft) whereas Rice (1967) reported a thickness of 8.7 m (26.4 ft).

Cameron (1918) first named the Fitzgerald Formation to describe the gypsum deposits exposed in the Gypsum Cliffs at Peace Point and the dolomitic limestone that is exposed along the Slave River and the escarpment west of Fort Smith. However, Cameron's (1918) usage now includes rocks of the Chinchaga Formation, Keg River Formation, Fort Vermilion Formation, and the Slave Point Formation.

The formation was redefined by Norris (1963) who restricted the occurrence of the formation (in Wood Buffalo National Park) to the type section. Norris (1963) described the formation as a pale to dark brown, slightly argillaceous, fine grained to granular, vuggy, locally fragmental and fossiliferous, laminated to thickly bedded cliff forming dolomite and dolomitic limestone. A thin bed of carbonaceous dolomite (2 in thick) is generally present near the base.

Norris (1963) suggested that the lower contact of the Fitzgerald Formation is gradational with the underlying La Loche Formation. However, Craig et al. (1967), argued that the formation was actually part of the Keg River Formation and that it unconformably overlay the La Loche Formation.

They suggested that the evaporites of the Chinchaga Formation were deposited around paleogeographic Precambrian highs. These highs had a local topography varying from 33 m (100 ft) to 98 m (300 ft, Craig et al., 1967). The fact that the Keg River Formation unconformably overlies the La Loche Formation suggests that the Chinchaga seas did not reach the higher parts of these topographic highs, or that later dissolution removed some of the evaporites.

The Fitzgerald Formation was originally assigned a late Silurian age by Cameron (1922) on the basis of paleontological work done by Kindle. Rice (1967) reported the occurrence of Welleria meadowlakensis in exposures along the Slave River. Rice (1967) compared the ostracod fauna from the Fitzgerald Formation with other Devonian faunas on a worldwide basis and suggested an Eifelian age for the formation, as well as suggesting that the ostracod Welleria meadowlakensis and the coral Planeophyllum planetum (Richmond, 1965) are "unique" to the Fitzgerald Formation. However, fauna does not define a lithostratigraphic unit; lithology does. Lithological descriptions of the Fitzgerald Formation (Norris, 1963) are similar to lithological descriptions of the Keg River Formation.

The Fitzgerald Formation was equated to the Keg River Formation by Craig et al. (1967) and is lithologically similar to that formation. These two points, along with the assumption that the formation exists only on the basis of faunal evidence, favours the view that some of the exposures

named the Fitzgerald Formation should be equated to the Keg River Formation of the eastern portion of Wood Buffalo National Park.

COLD LAKE FORMATION

The Cold Lake Formation is not easily identified in Wood Buffalo National Park. However, this formation does occur in boreholes adjacent to the south and east boundaries of the park where a maximum thickness of 88.6 m (270 ft) is present (Rice 1967). The Cold Lake Formation was reported in the Caribou Mountains, where 65 m (198 ft) occurs. The presence of brine salts with high Sodium and Calcium concentrations (over 90%) along the escarpment (Camsell, 1917) also suggests the occurrence of salt in the subsurface of Wood Buffalo National Park.

The Cold Lake Formation was named by Sherwin (1962) for a section composed predominately of salt. The type well is the E Cold Lake No. 7-22 Well (7-22-66-1W4) where the salt occurs at a depth of 929 m (2,832 ft; Rice, 1967).

The Cold Lake Formation comprises a clear salt that was dissolved and recrystallized prior to burial, thereby resulting in a low bromide content (Wardlaw and Watson, 1966). The lack of sulphate cycles in the salt sequences also suggests dissolution with subsequent recrystallization (Rice, 1967). Red dolomitic mudstone occurs throughout the formation in northcentral and northwestern Alberta and were originally thought to be a residue left by solution of the salt (Belyea, 1971). The red mudstone that forms marker bed

"H" at the top of the unit is thought to be a soil zone, thus representing a time of emergence over the Tathlina Arch (Belyea, 1971).

Direct evidence to suggest an unconformity at the top of the Cold Lake Formation in the study area is not available. However, as the Devonian strata wedges out over the Precambrian highs, erosion, solution, and/or non-deposition of the salt probably occurred. The fact that the Keg River Formation unconformably overlies the La Loche Formation (along the Slave River) suggests that there was dissolution, and/or non-deposition of both the Chinchaga Formation and Cold Lake Formation close to the edge of the Precambrian Shield. The upper contact is unconformable in northeastern Alberta because the lower member of the Chinchaga Formation was not deposited (Rice, 1967) . This is based on data from wells to the west and south of Wood Buffalo National Park, where the basal "detrital unit", which divides the Chinchaga into upper and lower members, occurs at the base of the Chinchaga Formation (Rice, 1967).

It would appear that where the Lower Chinchaga member is present, the Cold Lake Formation/Chinchaga Formation contact is conformable. However, where the Lower Chinchaga member is absent, as in northeastern Alberta, the upper contact of the Cold Lake Formation possibly represents an erosional surface. Following Rice (1967), the upper contact of the Cold Lake Formation, in the study area is considered to be an erosional surface that formed during Middle

Devonian (Eifelian) time.

THE CHINCHAGA FORMATION

The Chinchaga Formation was first defined by Law (1955) in the subsurface of central Alberta to describe a predominately anhydrite unit that contains minor amounts of dolomite. The lower boundary of the formation was lowered by Grayston et al. (1964) to include a lower detrital unit. Belyea and Norris (1962), Rice (1967) and Belyea (1971) divided the formation into lower and upper members. The lower member is predominately anhydrite. The upper member consists of two units, a basal unit of quartzose sandstone and a laterally equivalent bright green, locally red shale which grades upward into the anhydrite and fine grained dolomite of the upper member. The lower member occurs in the MacKenzie Basin to the north, but not in northeastern Alberta (Rice, 1967).

Belyea (1971) interpreted the basal sandstone of the upper member as channel fills that developed in association with the erosional surface at the top of the lower member. The contact between the two members is sharp with the underlying beds gradually truncated from west to east (Belyea (1971)). With work done in the area of the Tathlina Arch, Belyea (1971) showed this unconformity resulted from a period of uplift, regional tilting and erosion of the area.

Norris (1963) assigned the name Hay Camp Formation to a Group of rocks 3.2 km north of Hay Camp, on the western bank of the Slave River. The unit was described as being a

collapse breccia. In 1965, he lowered the rank of the unit to member status and placed it in the Chinchaga Formation. Craig et al. (1967) omitted the term Hay Camp Member from their nomenclature of the area, simply referring the breccia to the Chinchaga Formation. Due to the unmappable extent of the Hay Camp Breccia (on a 1:250,000 scale) the member status which was assigned by Norris (1965) is retained in this study.

Norris (1965) and Craig et al. (1967) placed the breccia, consisting of limestone and calcareous dolostone, in the Chinchaga Formation. Norris (1963) and Craig et al. (1967) placed the unconformity at the top of the Hay Camp Breccia, beneath the non-brecciated Keg River Formation. The extreme easterly location of that unconformity suggests that the Keg River Formation was deposited high on the Precambrian basement. Slight regressions of a Devonian Sea would have exposed the formation to surface conditions. Subsequent dissolution of the underlying evaporites would have caused collapse of the previously deposited Keg River Formation, thereby forming a breccia. This emergent surface and associated unconformity would be lost to the south and southwest where deeper water conditions and continued deposition occurred. However, low water conditions and possible emergent areas, over carbonate build-ups on the edge of the La Crete Basin are further represented by the breccias formed along the Salt River. Shallow water conditions and subaerially exposed surfaces in the Keg River

Formation in the Salt River study area probably coincide with the unconformity produced over the Precambrian highs to the east.

The upper contact of the Chinchaga Formation, at Hay Camp is indeterminable because the contact is lost within the breccia itself. However, west of the Precambrian high, the Chinchaga Formation is conformably overlain by the Keg River Formation with the unconformity occurring higher in the stratigraphic section.

The upper contact of the Chinchaga Formation was reported in the Salt River area by Tsui (1982) who defined it as the base of the stratigraphically lowest dark grey to black, fine grained to thin bedded limestone.

In Wood Buffalo National Park and just outside the park boundary the Lower Elk Point Sub-Group/Upper Elk Point Sub-Group contact occurs, in the Hay Camp Breccia, along the Salt River (north of the Salt River Bridge), at the Walk-in Cave (located 4.5 km northwest of the road to Pine Lake and 2 km northeast of the Parsons Lake Road).

C. UPPER ELK POINT SUB-GROUP

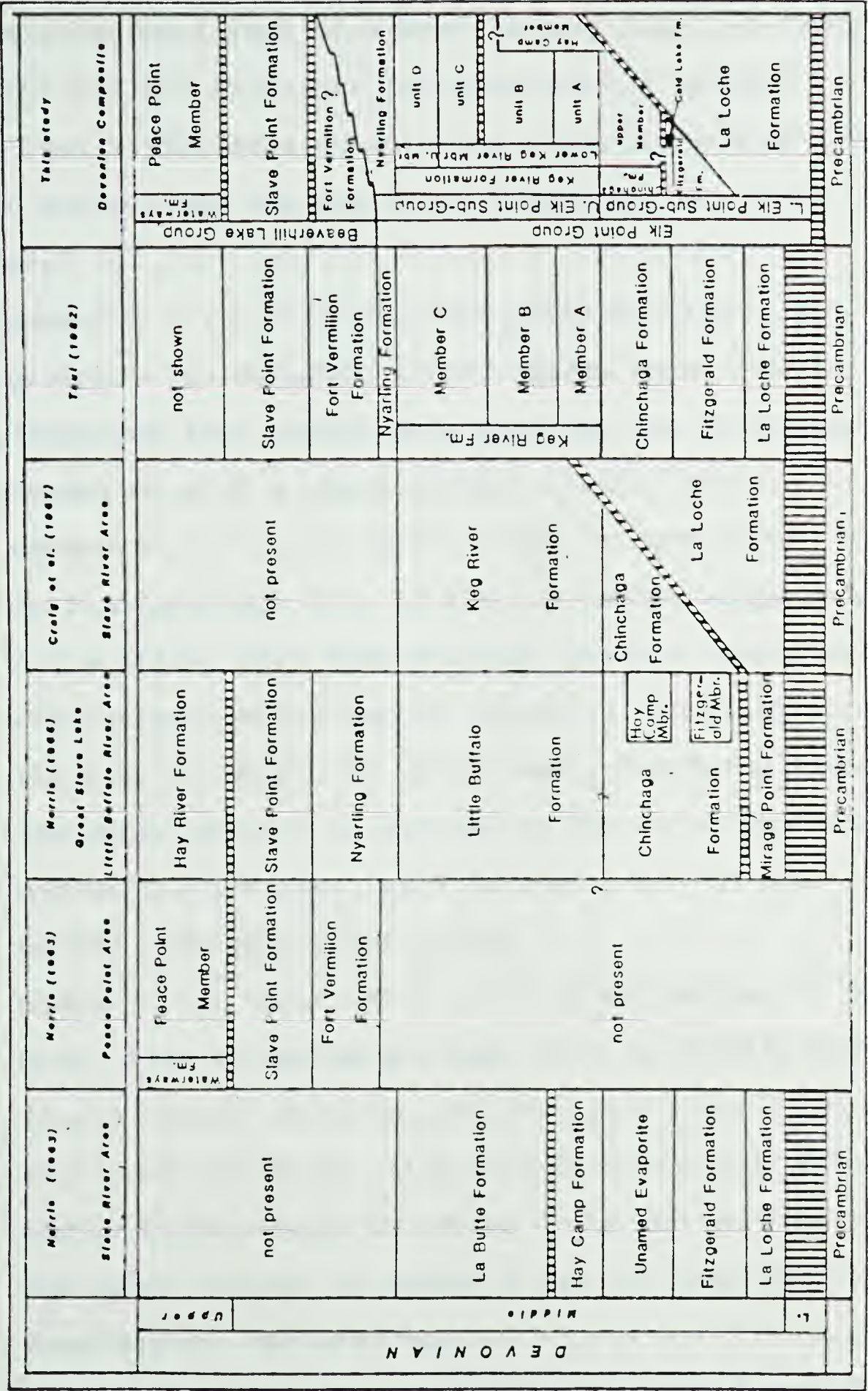
KEG RIVER FORMATION OF WOOD BUFFALO NATIONAL PARK

Based on paleontological work done by Kindle, Cameron (1922), assigned an Upper Silurian age to the evaporites and carbonates on Stoney and Caribou Islands (Slave River). He also assigned an Upper Silurian age to the evaporites and carbonates below the Main Falls on Little Buffalo River.

They were originally referred to the Fitzgerald Formation (Cameron, 1922). The limestones at La Butte and the Little Buffalo Falls were considered to be Devonian in age and therefore correlative with the Pine Point Formation to the northwest.

Norris (1963) termed the evaporites north of Hay Camp and the Escarpment west of Fitzgerald "the Unnamed Evaporites" (Fig. 6). The limestone breccia overlying the evaporites and unconformably overlain by the La Butte Formation was termed the Hay Camp Formation (Norris, 1963). The Unnamed Evaporites were redefined as the Chinchaga Formation with the Hay Camp breccia placed into this formation (Norris, 1965).

Craig et al. (1967) redefined the La Butte Formation as the Keg River Formation, based on the presence of an Atrypa Zone, which also occurs at the top of Norris's (1965) Little Buffalo Falls sequence (Keg River Formation equivalent). They also stated that the zone occurs west of La Butte and at Little Buffalo Falls. Ostracods from the limestone north of Hay Camp are equivalent to Braun's (1966) "b" and "c" microfaunal assemblages, considered diagnostic of the lower part of the Keg River Formation (Craig et al., 1967). Unfortunately, a biostratigraphic unit (such as the Atrypa Zone), according to the stratigraphic code, does not define a formation. Lithologically however, the unit at Hay Camp is identical to the Keg River Formation described elsewhere. Some of the "formations" that have been defined in Wood



Buffalo National Park have been largely based on faunal content and are therefore, of questionable status.

Tsui (1982) studied outcrops along and around the Salt River and divided the Keg River Formation into three members;

1. member A - a 2 m thick, dark green to black, very fine grained, laminated to thinly bedded fossiliferous limestone that conformably overlies the Chinchaga Formation with a sharp contact (Tsui, 1982).
2. member B - a 11.2 m thick, light to dark brown, coarse to fine grained, thin to thickly bedded vuggy dolostone. A 2 m thick, very fine grained, faintly laminated, argillaceous dolostone to dolomitic limestone occurs at the base of this unit, conformably overlying member A. The upper contact is defined by the first appearance of medium to dark grey, well jointed, thin to thickly bedded limestone (Tsui, 1982).
3. member C - a lower unit, 5.5 m thick, medium to dark grey, fine to medium grained, thin to thickly bedded fossiliferous limestone and an upper unit, 7.2 m thick, light grey to brown, fine to medium grained, thin to medium bedded vuggy dolostone to dolomitic limestone. The upper contact of member C has not been observed in Wood Buffalo National Park.

Tsui, (1982, Fig. 8, p.22) showed these 3 members to be correlatable across Wood Buffalo National Park.

KEG RIVER MARKER HORIZON

Corrigan (1975) described a regional marker horizon, which occurs in the La Crete Basin to the south of Wood Buffalo National Park. It is defined as a 5 foot (1.6 m) thick, black to dark brown, organic rich, laminated unit which exhibits a nodular, boudinaged, streaked out appearance with differential compaction occurring around the boudinage (Corrigan, 1975). Fauna of this unit includes Tasmanitids, Cricoconarids, rare crinoid ossicles and brachiopods (Corrigan, 1975). He thought that the unit marked a deepening of the sea in the La Crete Basin, thereby causing a change from oxidizing to reducing conditions on the sea floor.

Similar sedimentary units were reported by Wardlaw and Reinson (1971) from the lower Winnipegosis member of Saskatchewan (Corrigan, 1975). McCamis and Griffith (1967) described a thin, highly bituminous, Tentaculitid containing unit occurring at the top of the Lower Keg River Member in the Zama Basin. Tsui (1982) reported a nodular carbonate unit in the Firetower Doline (1.2 m thick) and the Walk-in Cave (1.4 m thick).

Corrigan (1975) stated that this zone possesses all the qualities of a marker horizon. He also argued that it probably approximates a time-stratigraphic unit and was thus used as a major datum in his study. Assuming the nodular carbonate unit reported by Tsui (1982) is the same lithologic unit as Corrigan's (1975) marker, it is also used

as a major datum in this study. In order to eliminate confusion with other marker horizons in the study area, this marker will be redefined herein as the Keg River marker.

TERMINOLOGY FOR THE KEG RIVER FORMATION OF NORTHERN ALBERTA

The different terminology used for describing the Keg River Formation of northern Alberta is confusing (Fig. 7). The Keg River Formation was originally defined by Law (1955). However, Hriskevich (1966), based on his work in the Rainbow Basin, divided the formation into a Lower Keg River Member, and an Upper Rainbow Reef Member. McCamis and Griffith (1967) adopted this two-fold division and divided the Keg River Formation of the Zama Basin into Hriskevich's Lower Keg River Member and their own defined Upper Keg River Reef Member. Their upper member only corresponds in part to Hriskevich's upper member.

In the La Crete Basin, Corrigan (1975) split the Keg River Formation into two members, the Lower Keg River Member, and the Upper Keg River Member. He further divided the Lower Keg River Member into the Lower Platform unit and the Upper Platform unit. Corrigan's (1975) Lower Keg River Member does not coincide stratigraphically with that defined by McCamis and Griffith (1967), but does include most of their Upper Keg River Member. The base of McCamis and Griffith's (1967) Upper Keg River Member is lithologically similar to Corrigan's (1975) Upper Platform division. If the 1.5 m (5 foot) Keg River marker is a time stratigraphic horizon, as Corrigan (1975) suggested, then his Upper

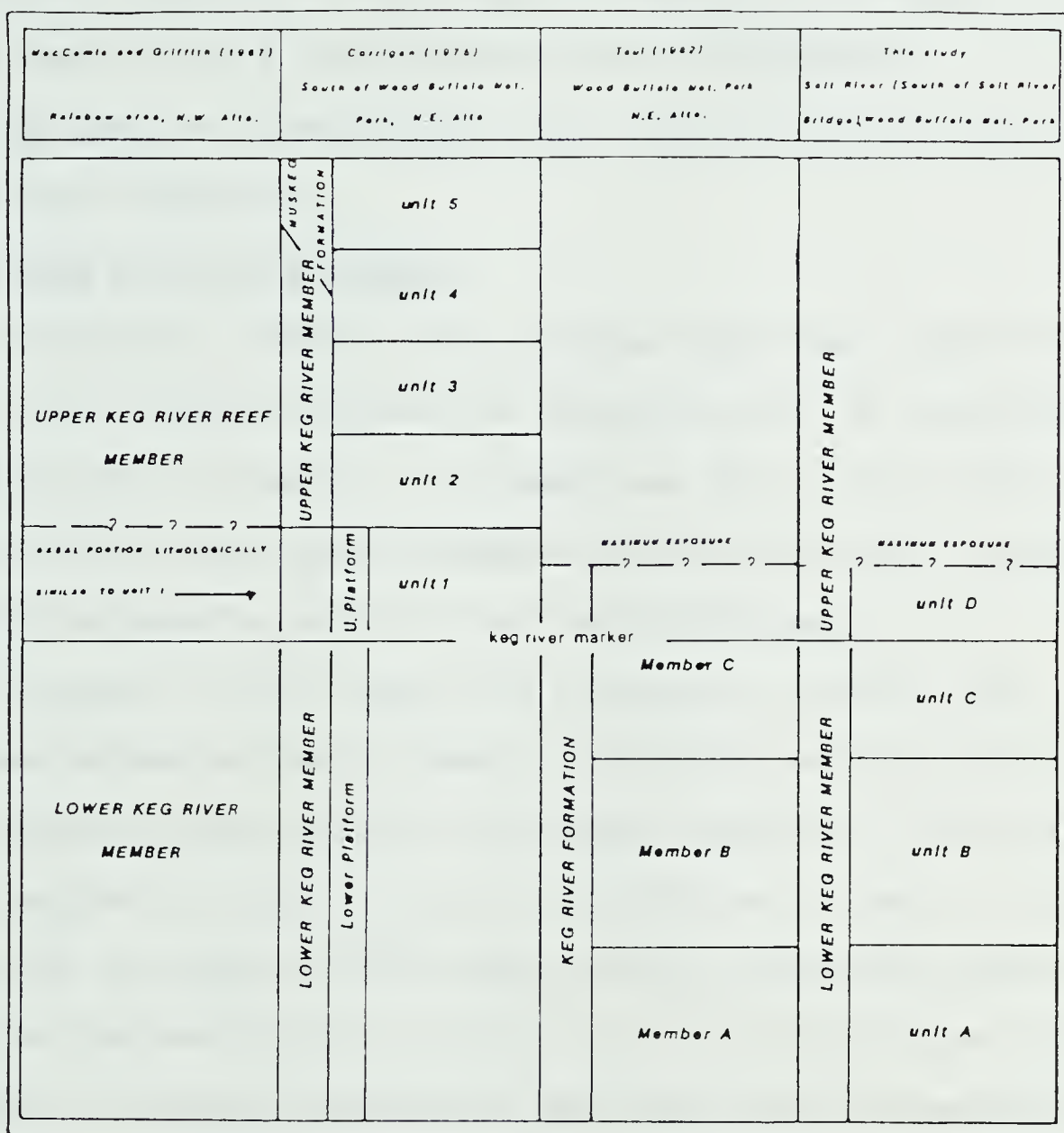


FIGURE 7. Nomenclature and Correlation of the Keg River Formation of Northern Alberta.

Platform division is stratigraphically equivalent to McCamis and Griffith's (1967) Upper Keg River Formation.

Tsui (1982), in defining his Members A through C, did not attempt to fit his members into the regional stratigraphy or terminology (other than to place them in the Keg River Formation).

THE LOWER KEG RIVER MEMBER

Hriskevich (1966) first defined the lower Keg River Member as a dense open marine carbonate unit at the base of the Keg River Formation. It comprises 49 to 53 m (150 to 160 ft) of dark brown, argillaceous calcilutite that contains crinoid columnals and thin-shelled brachiopods.

Corrigan (1975) described a massive, barren, buff, fine grained dolomite which commonly contained purplish blebs, conformably overlying the Chinchaga Formation. This unit is lithologically similar and correlatable with Tsui's (1982) member B. Corrigan (1975) reported that above the barren zone in the central area of the La Crete Basin there is a 16 m thick, strongly bioturbated, mottled mudstone/wackestone. Fauna is concentrated towards the top of the unit. Tsui (1982) also reported a lithologically similar unit with fauna that is concentrated toward the top of the unit. Crinoids, ostracods, rugose corals, and articulate brachiopods are the most common faunal elements. Corrigan (1975) stated that crinoids and articulate brachiopods form packstones near the top of the unit. This unit is correlatable with the lower portion of member C (below the

nodular carbonate unit) as defined by Tsui (1982).

The lower contact of the Lower Keg River Member is conformable with the underlying Chinchaga Formation. In the La Crete Basin the Member may grade up into the Keg River marker; however, a sharp boundary is more common (Corrigan, 1975).

UPPER KEG RIVER MEMBER

The Upper Keg River Reef Member consists of organic reefs (McCamis and Griffith, 1967). McCamis and Griffith (1967) showed that the lithologies vary according to their position with respect to the reef crest. Dolomitization of the reef is common. The basal 13 m (40 ft) of the Upper Keg River Reef Member consists of a dolomitized crinoidal calcarenite (McCamis and Griffith, 1967).

Corrigan (1975) described the 10 m thick upper platform unit (Unit 1) as an oncolitic floatstone with a mudstone/skeletal, predominantly wackestone matrix. A calcrete horizon occurs between Corrigan's (1975) upper platform unit 1 (regionally equivalent to the top of the lower Keg River Member) and Corrigan's (1975) upper platform unit 2 (regionally equivalent to the base of the Upper Keg River Member).

Tsui (1982) described the top portion of his member C, above the nodular carbonate unit (Keg River marker) as being dolomitic and vuggy. Neither calcrete nor oncolites were noted by Tsui (1982) in the upper 8.2 m of member C (above the nodular carbonate unit). However, reddish brown lumps of

clay (1/16" diameter) did occur (Tsui, pers. comm., 1983). Tsui (1982) described this portion of member C as a light to grey, fine grained, gritty, vuggy dolostone to dolomitic limestone that lacks fossils. The scarcity of fossils is due to more complete dolomitization of the Wood Buffalo outcrops. Ghosts, of fossils, and of the original limestone lithology are common evidence of extensive dolomitization.

UPPER KEG RIVER MEMBER OF THE LA CRETE BASIN

Corrigan (1975) defined five units (numbered 1 through 5) in the Keg River/Muskeg Formations. Corrigan (1975) provided detailed descriptions of the mounds, evaporitic mound caps, and intermound evaporites and carbonate laminates.

UNIT 1 - Unit 1 was the first unit deposited on top of the Keg River marker. North of the Peace River High the unit is a crinoidal rich, mound forming carbonate; whereas south of the Peace River High the unit is rich in stromatoporoids. Aveolitid corals and Amphipora are common. Corrigan (1975) reported breccias that formed as reefal debris on the slopes of the mounds. Unit 1 is a coarse pellet intraclast grainstone that contains atrypid brachiopods and Alveolites. Rocks from the northernmost Shell 10E Wood Buffalo Well (15-8-117-1W5) represent deposition on or close to the carbonate bank. There, unit 1 comprises grainstones of skeletal fragments, pellets, crinoids, corals and tabular stromatoporoids (Corrigan, 1975). Corrigan (1975, p. 155) stated "It seems that sedimentation on the bank along the

margin of the La Crete Basin was similar to the mounds in the basin."

To the south of the Peace River High, the top of unit 1 is well defined by a calcrete horizon. "...calcrete development includes solution pits, intense local brecciation, and the development of sharply undulatory surfaces" Corrigan (1975, p.100). Laminar crusts are well developed in some of the calcrete profiles. Generally, they occur beneath pisolitic zones.

On top of the Peace River High the top of unit 1 is marked by calcrete development and extreme vadose diagenesis (Corrigan, 1975). A brown dolomitization stage followed by a phase of white dolomitization has destroyed any evidence of primary fabric in some wells. Corrigan (1975) also suggested that the above replacement fabrics were followed by an anhydrite replacement phase.

The geologic history of the study area is marked by a number of major drops in sea level that caused large scale vadose diagenesis on the mounds, leaving behind evidence in the form of caverns, cavern collapse breccia, and weathered surfaces (Corrigan, 1975). The only extensive breccia that was reported from borehole data in Corrigan's (1975) study was from the Texaco 10E Wenzel (15-10-112-3W5). This breccia extends for 30.5 m below the top of unit 2. Clasts were generally elongate and oriented in all directions; but, the majority of the long axes of the clasts were vertical to subvertical. Corrigan (1975) noted that the carbonate in the

breccias is extensively dolomitized. Corrigan (1975, p. 155) stated that "The collapse was not caused by evaporite solution because the base of the zone is in the middle of Unit 1 crinoidal muds". Vadose solution and cavern collapse was caused by a drop in sea level and subsequent subaerial exposure that occurred at the end of unit 2 deposition and prior to unit 3 deposition (Corrigan, 1975). The collapse suggests a minimum drop in sea level at the end of Unit 2 deposition of over 30 m (Corrigan, 1975).

In the Mutex IOE Seaforth Well a 4.6 m cavern collapse was reported by Corrigan (1975). The collapse was thought to have been totally restricted to Unit 1 since no Unit 2 lithologies were observed in the breccia. Thus, Corrigan (1975) suggested that the breccia was caused by a drop in sea level, which occurred at the end of Unit 1 time. Extensive vadose diagenetic features were noted. Vadose channels acted as nuclei for diagenesis with white dolomite recrystallization and deposition being concentrated in and around them (Corrigan, 1975). In the Shell IOE Wood Buffalo Well (15-8-117-W5) an exposure surface, marked by channelling, is located midway through unit 3.

UNIT 2 - To the south of the Peace River High the unit is lithologically representative of a platform. It comprises carbonate mud, Amphipora and ostracods, and grades laterally into evaporite zones towards the basin margins. The unit only forms mounds to the northwest of the area. The unit is lithologically similar to Unit 1; however, in some wells,

the base of the unit is defined by the presence of oncolites. Intermound sediments of Unit 2 reflect shallow water facies throughout the La Crete basin. The top of the unit is well defined by a calcrete horizon and associated white dolomitization.

UNIT 3 - Unit 3 represents carbonate mounds which formed in the northern part of the La Crete Basin. To the South the unit consists of anhydrite and halite. Both mound and intermound lithofacies comprise barren carbonate laminates.

UNIT 4 - Unit 4 sedimentation also reflects mound growth in the northern part of the La Crete Basin. The unit comprises barren laminates that grade into anhydrite near the bottom and top of the mounds. The unit is overlain by halite everywhere in the basin.

UNIT 5 - Unit 5 is representative of a platform carbonate which occurs throughout the basin. It is separated from the underlying units by halite.

An important aspect of Corrigan's (1975) thesis is the interbedded nature of the mound sediments and the evaporites. As in the Rainbow-Zama Area, the evaporites have the potential of being dissolved. In the above basins, dissolution of the evaporites caused breccias and thicker accumulations of overlying rock (McCamis and Griffith, 1967).

THE NYARLING FORMATION

The Nyarling Formation was originally defined by Norris (1965) in a northwesterly trending belt that occurs between the northern boundary of Alberta and 16 km south of Great Slave Lake. The formation consists of gypsum with minor amounts of brown, thinly bedded, fissile, fine grained to aphanitic limestone, which contains dark brown carbonaceous streaks (Norris, 1965). The unit outcrops around the edges of collapse sinkholes that are located throughout Wood Buffalo National Park (Airphoto Analysis Associates Consultants Ltd., 1978). Neither the upper nor the lower contacts are exposed in the study area.

The Nyarling Formation is possibly the stratigraphic equivalent of the upper 80% of the Pine Point Formation, the combined Muskeg Formation and Watt Mountain Formation (as defined by Law, 1955), and the Fort Vermilion Formation (Norris, 1965). Based on an assumed southwest regional dip of 13 ft per mile, for the northern area of Wood Buffalo National Park, Norris (1965) calculated a minimum 138 m (420 ft) thickness for the Nyarling Formation. It is important to realize that the Muskeg Formation is stratigraphically equivalent to the Nyarling Formation, for the former is found flanking, and forming between reefs, in the Rainbow-Zama Basins to the west and the center of the La Crete Basin to the south.

D. BEAVERHILL LAKE GROUP

The Beaverhill Lake Group encompasses evaporites, limestones and shales which occur between the Middle Devonian Elk Point Group and the Upper Devonian Woodbend Shales of central Alberta. The upper boundary of the Beaverhill Lake Group occurs at the top of the Waterways Formation. The basal boundary is not so well defined. Fong (1959, 1960), Edie (1961) and Jenik and Lerbekmo (1968) placed the base of the Beaverhill Lake Group at the top of the Gilwood Sandstone Member of the Watt Mountain Formation. However, Thomas and Rhodes (1961) and Murray (1965) placed the boundary at the top of the Fort Vermilion Formation. Leavitt and Fischbuch (1968) placed the base of the Beaverhill Lake Group at the base of the Fort Vermilion Formation and the top at the base of the Woodbend Shales. Leavitt and Fischbuch's (1968) nomenclature was used in this study simply because of its easier application to the study area (Fig. 8).

The Beaverhill Lake Group was first assigned the name Waterways Formation by Warren (1933) following work done in northeastern Alberta. The Waterways Formation was later divided into (in ascending order): the Firebag Member, the Calument Member, the Christina Member, the Moberly Member, and the Mildred Member (Crickmay, 1957). Norris (1963) renamed the basal 1.8 m (5.5 ft) of the Waterways Formation as the Livock River Formation.

The Geological staff of Imperial Oil Limited (Imperial Oil Ltd., 1950) termed an alternating 237 m (722 ft) thick sequence of limestone and shales in central Alberta the Beaverhill Lake Group. Leavitt and Fischbuch (1968) included the Fort Vermilion Formation, the Swan Hills Formation and the Waterways Formation in the Beaverhill Lake Group.

The Swan Hills Member of central Alberta was defined and divided into two units (Fong, 1959, 1960); an upper light brown unit and a lower dark brown unit. His Upper Beaverhill Lake unit was designated the Waterways Formation (Murray, 1965) and the basal Beaverhill Lake unit was given formational status based on work done in the Gypsum Cliffs area (Norris, 1963). Murray (1966), because of the Swan Hills Member's broad areal extent, raised it to formation status.

The stratigraphic relationship between the Slave Point and Swan Hills formations of central Alberta is obscure. In the area just west of the Peace River Arch the two formations are lithologically identical and probably represent the same formation. The Swan Hills Formation represents a diachronous equivalent of the Slave Point Formation and occurs in a higher stratigraphic position due to onlap over the Peace River Arch. Depending on the proximity to the Peace River Arch, the Slave Point Formation may have acted as a platform on which the highly fossiliferous Swan Hills strata accumulated or the two Formations may represent the same diachronous unit. However,

		Upper Devonian		MIDDLE DEVONIAN									
Warren, 1933 (N.E. Alta)	Waterways Formation										Elk Point Group		
Geological Staff Imperial Oil, 1950 (Edmonton Area)	Beaverhill Lake Formation										Elk Point Group		
Crickmay, 1957 Norris, 1963 (N.E. Alta)	Waterways Formation					Livock River Formation					Elk Point Group		
	Mildred Member	Moberly Member	Christine Member	Celumet Member	Firebag Member								
Fong, 1959, 1960 (Sven Hills Area)	Beaverhill Lake Formation					Sven Hills Member					Basal Beaverhill Lake	Elk Point Group	
	Upper Beaverhill Lake					Light Brown Unit	Dark Brown Unit						
Murray, 1966 (Judy Creek Area)						Waterways Formation					Fort Vermillion Formation	Elk Point Group	
Leavitt and Fischbuch, 1968 (Sven Hills Area)	Beaverhill Lake Group					Swan HILLS					Fort Vermillion Formation	Elk Point Group	
	Waterways Formation												
This Study Wood Buffalo National Park (Peace Pt. Area)	Beaverhill Lake Group					Slave Point Formation					Fort Vermillion Formation	Elk Point Group	
	Peace Point Member	Waterways Formation											

FIGURE 8. Devonian Nomenclature and Correlation of the Beaverhill Lake Group of Alberta.

the Swan Hills/Slave Point formations of central Alberta and the Slave Point Formation of Wood Buffalo National Park are lithologically different, therefore being properly defined as distinct, separate, lithostratigraphic units. Since they are lateral equivalents, the Slave Point Formation of northeastern Alberta has been placed in the Beaverhill Lake Group (Fig. 8).

In Wood Buffalo National Park the Fort Vermilion Formation and the Waterways Formation are lithologically similar to the type well and type section.

FORT VERMILION FORMATION

Evaporites of the Gypsum Cliffs area were raised to formation status by Norris (1963). "The Fort Vermilion Formation consists mainly of thin - to thick - evenly bedded white gypsum with thin interbeds and laminae of varicolored, commonly bluish grey and light brown impure gypsum."

(Norris, 1963, p. 61). Local folding, faulting and movement along joint planes made it difficult to obtain a reliable estimate for the thickness of the formation, but Norris (1963) suggested that the formation has a maximum thickness of 24.6 m (75 ft). In the Swan Hills type well (Home Regent A 10-10-67-10W5M) the formation is 8.2 m (25 ft) thick. The lower contact of the formation was not observed at Peace Point. In the type well, the shales and sandstones of the Watt Mountain Formation (Lower Elk Point Sub-Group) underlay the Fort Vermilion Formation.

It is possible that the evaporites exposed in the Gypsum cliffs area are not representative of the Fort Vermilion Formation. The presence of an angular unconformity between the evaporites and the Slave Point Formation (this thesis) and a slight lithological discrepancy with the Fort Vermilion Formation of north-central and central Alberta (Richmond, pers. comm., 1984) suggests that the evaporites are possibly representative of another formation (possibly the Muskeg or Nyarling Formations). However, the lack of substantial data and evidence, precludes a definite conclusion in this respect.

UPPER CONTACT OF THE FORT VERMILION FORMATION IN WOOD BUFFALO NATIONAL PARK

Crickmay (1957) stated that an angular discordance exists between the Fort Vermilion Formation and the Slave Point Formation. Norris (1963, p. 61), however, argued that the contact is conformable, explaining the discordance as "...the result of local differential deformation of mobile evaporites and the more competent carbonates of the Slave Point Formation." Norris and Uyeno (1983, p. 7) further stated "In addition, lithological similarities in the boundary beds between the Fort Vermilion and Slave Point formations strongly suggest a transitional boundary rather than a hiatus between the two rock units. On this evidence, the units are interpreted by Norris as being conformable." Although the Fort Vermilion gypsum had been extremely mobile, the Fort Vermilion/Slave Point contact observed in

the brecciated zone along the northern cliffs of the Peace River is an unconformity (this thesis).

SLAVE POINT FORMATION

The Slave Point Formation was defined by Cameron (1918) following his work in the Great Slave Lake area. Later, Norris (1963) assigned rocks that outcrop along the Peace River (around Boyer Rapids) to the Slave Point Formation. Based on fossil content, he assigned them a Middle Devonian age. Kindle (1928) had previously assigned a Silurian age to this formation.

Norris (1963) referred to the Slave Point Formation as limestones and dolomitic limestones overlying the Fort Vermilion evaporites and underlying bluish green shales of the Waterways Formation. Norris measured 16 to 23 m (50 to 70 ft) of section. He stated that the varying thicknesses suggested a pre-Waterways erosion surface. In the Zama region, the varying thickness of the formation is due to removal of the Black Creek Salt (Lower Muskeg; McCamis and Griffith, 1967). Both suggestions are valid and both probably account for the varying thicknesses of the Slave Point Formation.

UPPER CONTACT OF THE SLAVE POINT FORMATION IN WOOD BUFFALO NATIONAL PARK

The upper contact of the Slave Point Formation was not observed in the study area. However, as will be shown, the top of the exposed section, around Peace Point, does suggest a pre-Waterways erosional surface. Norris (1963) reported

that beds of the Peace Point Member lie in depressions in the Slave Point Formation. These depressions are suggestive of sinkholes and widened joint fissures that are typical of a karst topography. Norris (1963) and Norris and Uyeno (1983) suggested that some of the brecciation is of a solution collapse origin. "The karst topography within the Slave Point beds was probably initiated during this hiatus before Peace Point beds were deposited. Undoubtedly some solution of the Fort Vermilion evaporites and collapse of the Slave Point beds have occurred since the Peace Point beds were deposited because the latter are also brecciated in some places" (Norris, 1963, p. 67). The sharp change in lithology, as well as a profound faunal break between the Slave Point Formation and the Waterways Formation (Peace Point Member), in the Gypsum Cliffs area, is strong evidence that a depositional hiatus occurred between the two formations (Norris, 1963; Norris and Uyeno, 1983). However, the hiatus was of short duration because there are no missing conodont zones across the Slave Point Formation/Waterways Formation boundary (Norris and Uyeno, 1983). Fauna of the Peace Point Member correlates with the Polygnathus asymmetricus Zone of the Firebag Member of northeastern Alberta (Norris and Uyeno, 1983). "It is the base of this interval, marked by the lowermost P. asymmetricus Zone, that most Devonian workers in Western Canada have tentatively and traditionally selected as the base of the Upper Devonian". Norris and Uyeno (1983) further

stated that the faunal break recognized between the Slave Point Formation and the Waterways Formation (Peace Point Member) in the Gypsum Cliffs area may be the same break recognized elsewhere by Norris (1965), McLaren in Belyea and McLaren (1962), Fischbuch (1968), McGill (1963, 1966), Braun (1968, 1977), Norris et al. (in press), Pedder (1975), and Uyeno (1978).

WATERWAYS FORMATION

The Waterways Formation was defined by Warren (1933) for rocks that are overlain by Lower Cretaceous strata in the Fort McMurray area. However, in the Gypsum Cliffs area, Kindle (1928) proposed the name "Peace Point Beds" for a fossiliferous blue shale and thin bedded limestone. Based on priority, Norris (1963) retained Kindle's (1928) terminology, but assigned a member status to the Peace Point beds. The "Peace Point Member" is also retained for this study. Norris (1963, p. 65) described the Peace Point Member as "a light to dark olive green soft shale, with thin rubbly interbeds of argillaceous limestone." The Peace Point Member attains a maximum thickness of 6.1 m in the Gypsum Cliffs area (Norris and Uyeno, 1983). On the south bank of the Peace River, 2.3 km downstream from the lower end of Boyer Rapids, the lowest 2 m (6 ft) of the Peace Point Member consists of "light brownish, thin bedded rubbly limestone, dark brownish grey limestone, and dark brownish grey carbonaceous shale; interlaminated with dark brown argillaceous limestone" (Norris, 1963, p. 65). Beds of the

Peace Point Member are unconformably overlain by Pleistocene and Recent deposits in the Gypsum Cliffs area (Norris, 1963; Norris and Uyeno, 1983).

In the northern portion of Wood Buffalo National Park the Waterways Formation is referred to as the "Hay River Formation". There, the Hay River Formation comprises a greenish grey shale, limestone, and sandstone (Air Photo Analysis Associates Consultants Limited, 1978). Bayrock and Reimchen (1976) also suggested that the Hay River Formation that occurs north of the park is equivalent to the Waterways Formation that occurs in the park. However, Belyea and McLaren (1962) suggested that the Waterways Formation correlates approximately with the lower half of the lower unnamed shale member of the Hay River Formation in the southern District of Mackenzie.

E. LOWER CRETACEOUS

Little work has been done on the erosional remnants of Cretaceous Strata that outcrop in Wood Buffalo National Park. However, in a composite section of Wood Buffalo National Park, Bayrock and Reimchen (1976, p. 5), included six Lower Cretaceous formations. They stated that the Cretaceous part of the composite section was not based solely on Cretaceous rocks that outcropped in the park, but on Cretaceous erosional remnants that occurred south of the park boundary in the Birch Mountains. They suggested that more than 610 m (2,000 ft) of Lower Cretaceous strata occur

in the Birch Mountains.

The Lower Cretaceous formations that occur in the Birch Mountains and surrounding Fort McMurray-Athabasca areas are, in ascending order; the McMurray Formation (including 4 informal units), the Clearwater Formation, the Grand Rapids Formation, the Joli Fou Formation, the Pelican Formation, and the La Biche Formation. These formations have been described by various authors, but the following descriptions are based primarily on work done by Carrigy (1959, 1973).

McMURRAY FORMATION

The name McMurray Formation was proposed by McLearn (1917) for oil bearing sandstones that unconformably overlie the Upper Devonian limestones of the area. He placed the top of the unit at the base of a green, argillaceous sandstone of the Clearwater Formation. The formation ranges in thickness from 30 m to 61 m (Bayrock and Reimchen, 1976).

Carrigy (1973) stated that the McMurray Formation could be divided into four informal stratigraphic units; namely:

1. Pre-McMurray Beds Carrigy (1973, p. 80) defined these beds as "Coarse grained, quartzose sandstone, connected by silica and goethite, which appear to unconformably underlie the McMurray Formation."
2. Lower Unit of the McMurray Formation Carrigy (1973, p. 81) described the unit as "Lenticular beds of conglomerate, sand, shale, and silt which occupy the deeper depressions on the Pre-Cretaceous erosion surface." The sands contain wood, angular to well

rounded quartz grains, feldspars and mica, boulders of Athabasca sandstone, angular iron quartz, and some white quartzite grains. Individual grains show fracture, crystal intergrowths, granoblastic texture, and strain extinction. These sands are interbedded with grey micaceous siltstones. The top of the unit is a black to dark grey carbonaceous shale that contains marcasite cement and marcasite concretions (Carrigy, 1959). Basal strata of the unit consists of residual clays that are thought to be derived from the Waterways limestone.

3. Middle Unit of the McMurray Formation Carrigy (1973) described the unit as a muscovite containing, oil cemented quartz sand. These sands are interbedded with lenticular beds of micaceous silts, shales, and clays. He described the unit as being characterized by current bedding, plant remains, worm casts, logs of wood, and thin coal beds.
4. Upper Unit of the McMurray Formation The upper unit is lithologically similar to the middle unit. "Large shallow channels or scours filled with silt beds and siderite-cemented siltstones are present (Carrigy, 1973, p. 81)." The upper unit is differentiated from the middle unit by brackish-water fauna which is contained in the former unit (Mellon and Wall, 1956).

The Cretaceous outlier, west of Pine Lake, in Wood Buffalo National Park, is questionably named as the McMurray Formation on the C.S.P.G geological highway map

of Alberta (Canadian Society of Petroleum Geologists, geological highway map series, second edition, 1981).

CLEARWATER FORMATION

McConnell (1893) described the "Clearwater Shale" that outcropped below Grand Rapids, at Point La Biche, on the Athabasca River. McLearn (1917) described the Clearwater Formation as "soft grey shales, black shales, grey and green sandstone with black shale that contains poorly preserved Inoceramus."

Stelck et al. (1956), based on Foraminifera, assigned a middle Albian age to the formation.

GRAND RAPIDS FORMATION

The Grand Rapids Formation was first defined by McConnell (1893). The formation outcrops in the Athabasca River Valley west of Fort McMurray, on the Clearwater River and on the eastern side of Birch Mountain.

The Grand Rapids Formation is a 91.4 m thick, poorly cemented "salt and pepper sand". It contains quartz, feldspar, glauconite, chert, muscovite, and halite (Carrigy, 1959). Spherical calcareous nodules, reaching 10 ft in diameter are concentrated near the base of the formation.

JOLI FOU FORMATION

The unit was first named the Pelican Shale (McConnell, 1893). Wickenden (1949) renamed the formation the Joli Fou Formation. He applied the name to a 33.5 m thick dark grey marine shale which outcrops near the Joli Fou Rapids on the Athabasca River.

Stelck et al. (1956), studying Foraminifera, equated them with the Haplophragmoides gigas zone of middle Albian age.

PELICAN FORMATION

On the Athabasca River, southwest of Fort McMurray, the formation comprises a series of sandstone and siltstone lenses that are of a marine origin (Carrigy, 1973).

LA BICHE FORMATION

Grey shales, containing limestone concretions, that outcrop 3.2 km below Moose Lake were assigned to this formation by McConnell (1893). Bayrock and Reimchen (1976) described the unit as a very soft unconsolidated marine shale. Strata which overlie the Pelican Formation in the Birch Mountains area have not yet been studied (Carrigy, 1973).

III. STRATIGRAPHY OF THE KEG RIVER FORMATION, SALT RIVER AREA

A. THE KEG RIVER MARKER OF WOOD BUFFALO NATIONAL PARK

In Wood Buffalo National Park, Tsui (1982) reported a nodular carbonate unit in exposures at the Fire Tower Doline and the Walk-in Cave where thicknesses of 1.2 and 1.4 m occur respectively. This unit appears identical to the Keg River marker reported to the west and southeast of Wood Buffalo National Park.

In Wood Buffalo National Park the marker comprises a boudinaged, black to dark brown, nodular, fossiliferous, finely crystalline limestone. The elliptical boudins, which taper laterally, are up to 3 cm long. They are highly recrystallized but ghost structures of brachiopod spines and tentaculitids occur. Recrystallization gives a peloid, intraclastic appearance to the boudins (Figs. 9A, 9B).

The groundmass around the boudins is a highly argillaceous, finely crystalline limestone. The matrix, which forms up to 50% of the unit, contains 20% tentaculitids and/or brachiopod spines, 10 to 25% sponge spicules and indeterminate bioclastic debris, and less than 10% disarticulated thinly shelled brachiopods (Figs. 9C, 9D). The matrix is not as aggraded as the boudins, however, intense recrystallization does occur along planes that are concentrated with allochems. The greater the percentage of allochems, the greater the recrystallization. In some

laminae, concentrated areas of allochems cause small irregular shaped nodules which may represent the initial stage of development of the boudins. The boudins may be the result of recrystallization of the original limestone. Throughout the marker, recrystallization occurs on the allochem edges and has worked its way into the center of the allochem and away from the outer edge of the allochem into the matrix (Figs. 9A, 9B). Eventually, these aggradation fronts coalesce to form, what appear to look like a boudin. Compaction by overlying strata probably accentuates the shape of the boudin. Opaques, allochems and other material in the matrix bend around and are aligned parallel to the outer edges of the elongated boudins (Fig. 9D). The matrix has undergone various degrees of aggradation and in places relict patches of original micrite give the matrix a peloidal texture. Stylolites commonly occur between allochems that are in contact with one another.

B. THE KEG RIVER UNITS, SALT RIVER AREA

INTRODUCTION

Tsui (1982) defined three informal members in the Keg River Formation and demonstrated that they are correlatable throughout Wood Buffalo National Park. The three members defined by Tsui (1982) are now, for the most part, equated to the Lower Keg River Member as defined by Hriskevich (1966), based on similar lithologies and relationships to the Keg River marker (Fig. 7). According to the code of

FIGURE 9

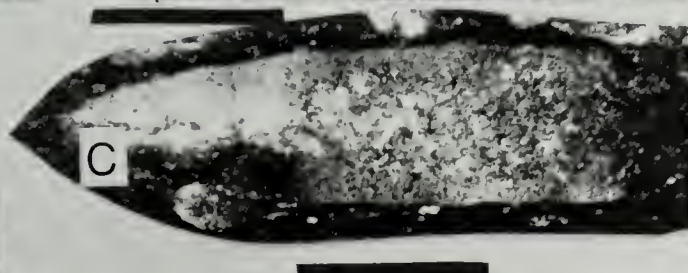
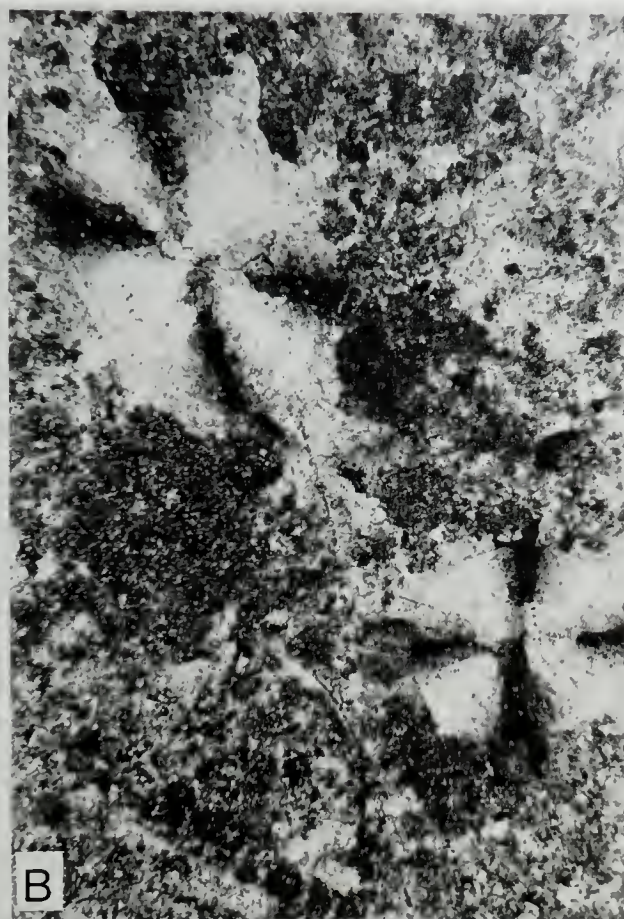
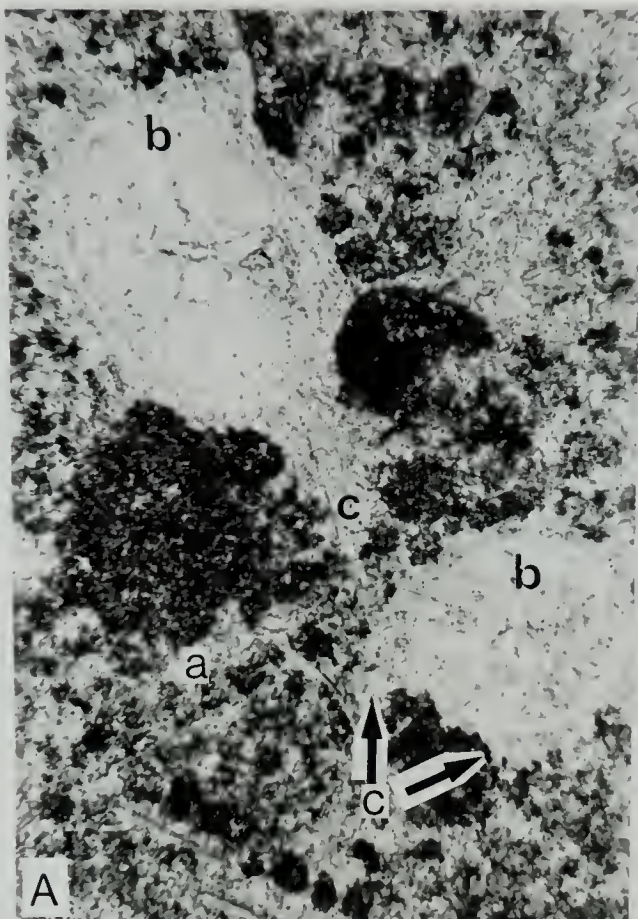
9A) Photomicrograph of a boudin in the Keg River Marker. Boudin displays a clotted texture (a). Ellipsoidal allochems (b) are either cross-sectional views of brachiopod spines or Tentaculites. Aggradation of the calcite in and around these allochems is evident with aggradation fronts moving into the centre of the allochems and out into the matrix (c), from the rim of the allochem. Thin section under plane-polarized light. Scale bar equals 0.4 mm.

9B) Photomicrograph same as 9A, except thin section is under cross-polarized light. Allochems display a pseudo-uniaxial cross. Scale bar equals 0.4 mm.

9C) Photomicrograph of longitudinal section of Tentaculites. Thin section under plane-polarized light. Scale bar equals 0.4 mm.

9D) Photomicrograph of the argillaceous matrix of the Keg River marker. Tentaculites and brachiopod spines are abundant. Thin section under cross-polarized light. Scale bar equals 0.4 mm.

9E) Field photograph of units A to C. The unit B3/unit C contact represents a diastem that divides dolostone from an overlying limestone. The break is also defined by the mottled texture of unit C. Note the wavy, stromatolitic-like bedding of unit B2. Meter stick for scale.



unit C,

unit B3

unit B2

unit B1

unit A

stratigraphic nomenclature (American Commission on Stratigraphic Nomenclature, article 7c, p.7, 1970) "Members may contain beds but never members of members". Therefore, the 3 members defined by Tsui (1982) in Wood Buffalo National Park cannot be designated members because they belong to both the Lower and Upper Keg River Members, as defined on a regional scale. Therefore, members A through C are herein referred to as units A through C. However, an additional complication is the fact that the Keg River marker occurs in the middle of Tsui's (1982) member C. This implies, therefore, that the rocks above the marker are the basal units of the formally designated Upper Keg River Member. Therefore, the uppermost 8.2 m of member C, as defined by Tsui (1982), is herein named unit D.

Based on the new terminology, units A through C belong to the Lower Keg River Member whereas unit D, which occurs above the Keg River marker, belongs to the Upper Keg River Member.

The thickness of strata between the base of the Keg River Formation and the Keg River marker is 18.6 m in the Walk-in Cave (Tsui, 1982), 16 m in the La Crete Basin (Corrigan, 1975) and 16.6 m south of the Salt River Bridge (Appendix 1, 2, this thesis). Along the Salt River 14.5 m of unit C overlies 1.35 m of unit B, which overlies 0.75 m of unit A. The Keg River marker should have occurred in the 14.75 m thick deposit of unit C along Salt River, but it did not. This implies that unit C, measured along the Salt

River, is at least 2 m thicker than unit C, 14 km to the southwest in the Walk-in Cave. This extra section, along with an abundant and highly diverse fauna, suggests the presence of a small carbonate build-up (A carbonate build-up refers to "a body of locally formed [laterally restricted] carbonate sediment which possesses topographic relief. This term carries no inference about internal composition" [Wilson, 1975].) in the platform (Carbonate platform refers to "huge carbonate bodies built up with a more or less horizontal top and abrupt shelf margins, where high energy sediments occur" [Wilson, 1975].) itself. Localized concentrations and patches of Alveolites and associated fauna suggest that reefoid and mound initiating processes occurred along the edge of the La Crete Basin. Lateral restriction of the Salt River carbonate buildup is not evident.

The extra thickness in the section at South Salt River Bridge area may simply be an expected northwesterly thickening of the platform. The platform thickens considerably to the north (Rainbow-Zama area) from the center of the La Crete Basin. The Lower Keg River Member is 16 m thick in the La Crete Basin (Corrigan, 1975), 18.6 m thick at the Walk-in Cave (Tsui, 1982) and 53 m thick in the Rainbow - Zama Area (McCamis and Griffith, 1967).

UNIT A

Unit A is the basal unit in the Salt River sections. It is exposed at the base of section H and also occurs in the

brecciated sections F, G and I. The lower contact is covered while the upper contact is abrupt and planar. Unit A is a recessive, medium grey to brown, thinly laminated to thinly bedded, dolomitic aphanocrystalline limestone that contains dark brown 2 micron wide, wispy, subhorizontal to horizontal, argillaceous laminae (Fig. 10A). The unit breaks easily along the thin argillaceous, organic laminations. Unit A has a peloidal appearance, however the peloid texture is due to recrystallization (aggradation) of an original finer grained micrite. The rock consists of 45 to 50% of < 4 micron wide micrite and 25 to 30% of 160 micron diameter neomorphic spar. Dolomite, which forms up to 10% of the rock, occurs in two distinct modes: an anhedral to euhedral 2.5 micron wide dolomite which occurs in amoeboid clusters and a coarser, 44 micron wide, white euhedral dolomite. A fibrous, twinned, indistinct calcite cement forms around allochems and increases in size away from the allochem surface. The unit has 5 to 10% intercrystalline porosity that increases towards the top of the unit. Unit A is partially filled with less than 5% bitumen and organics.

UNIT B

Unit B consists of three subunits that are exposed in section H, in part in section J and in the breccias of sections G and I.

UNIT B1

Unit B1 overlies unit A with a sharp and planar contact, whereas the upper contact is gradational over

FIGURE 10

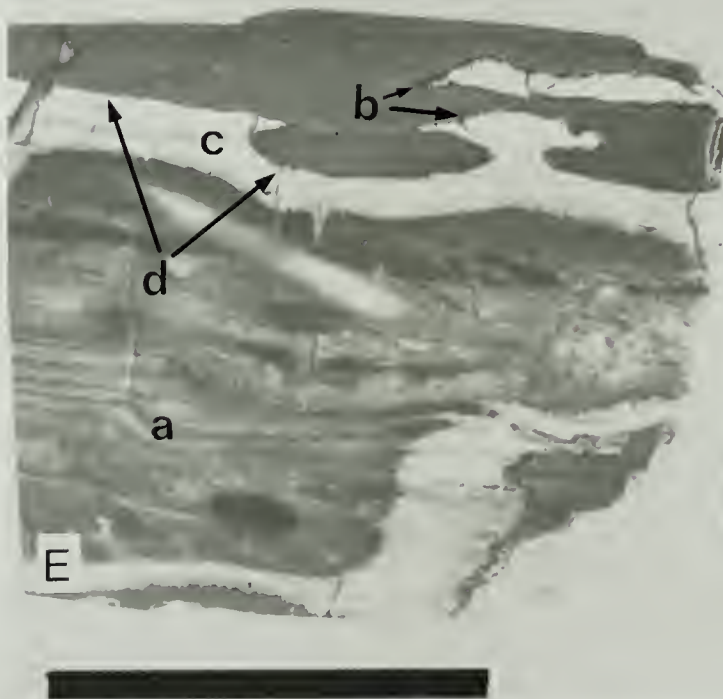
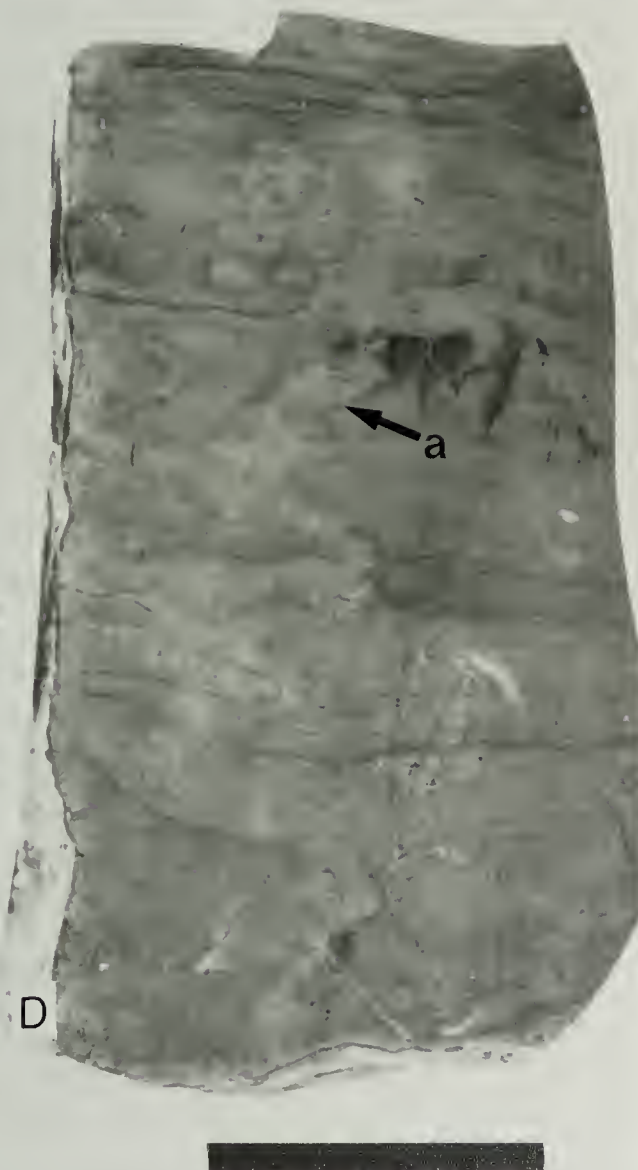
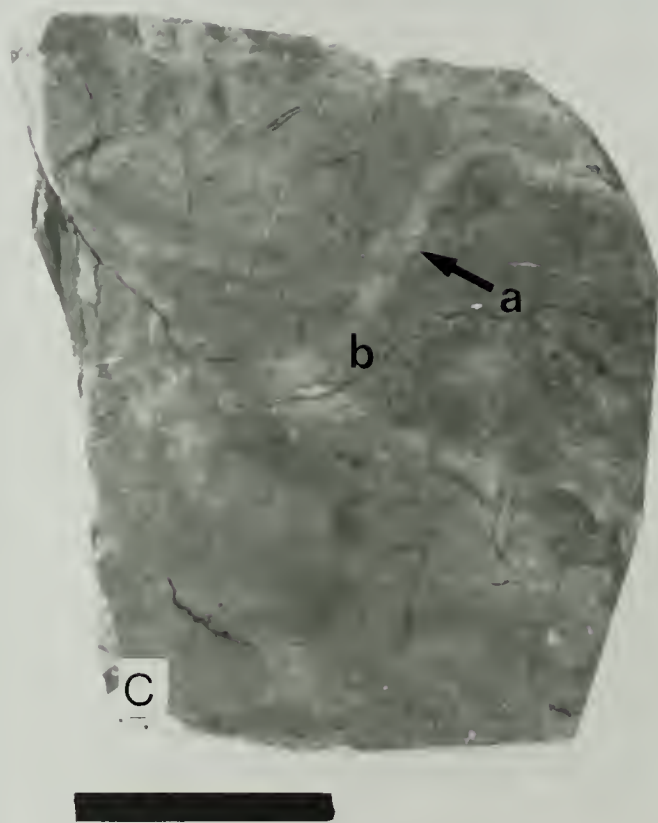
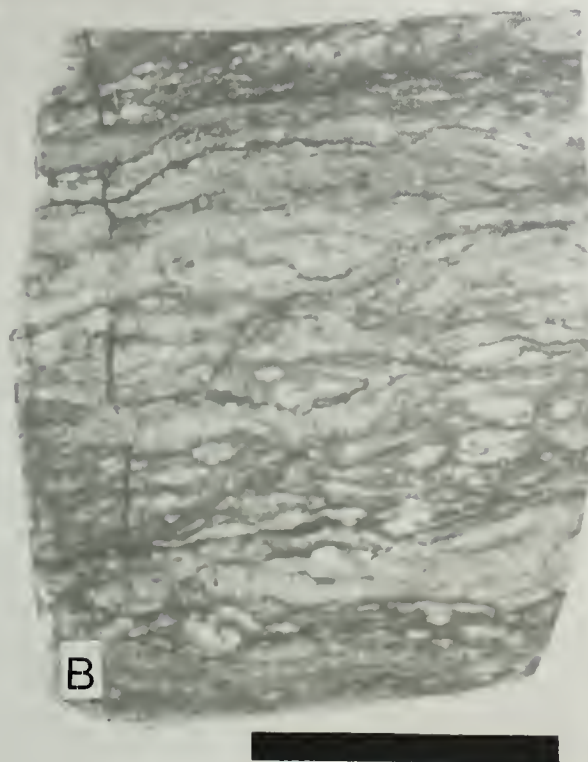
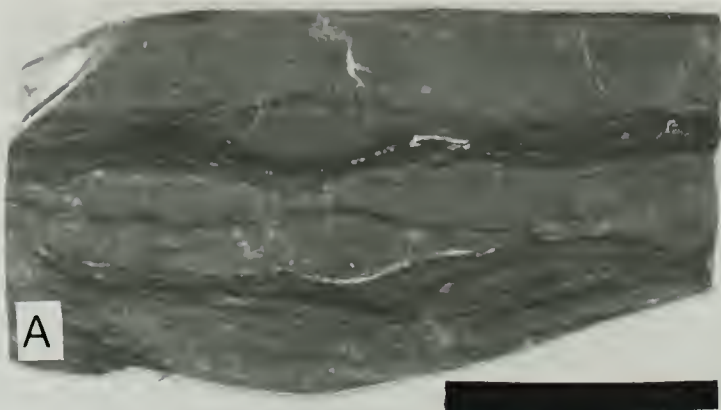
10A) Hand specimen of thinly laminated aphanocrystalline limestone from unit A in section A. Scale bar equals 3 cm.

10B) Hand specimen of calcareous, finely crystalline dolostone from unit B1 in section H showing the nodular, rubbly texture and laterally discontinuous, argillaceous laminae. Scale bar equals 3 cm.

10C) Hand specimen of calcareous, very finely crystalline dolostone from unit B2 in section H. Lines of dissolution (a) pervade throughout the unit with fractures abruptly ending at the line of dissolution (b). Scale bar equals 3 cm

10D) Hand specimen of calcareous, very finely crystalline dolostone from unit B2 in section J showing the thin laterally discontinuous, argillaceous laminae. The overall texture is similar to a "patterned dolostone" (a). Scale bar equals 3 cm.

10E) Hand specimen of a calcareous to cherty, very finely crystalline dolostone from unit B2. Sample taken from talus at base of pseudoanticline in section A. Dark areas are completely chert replaced (a). Opaque lined surfaces occur (b) and commonly truncate fractures (d). Laminations are continuous across the opaque lined surfaces (c), thus suggesting that the surfaces are not hardgrounds. Note the textural/lithological change in unit B2 from different sections. Scale bar equals 3 cm.



4 cm. In section H, unit B1 forms a wedge that thins towards section I. This unit comprises a semirecessive, orangey buff to cream colored, very thinly bedded to thinly bedded, calcareous, finely crystalline dolostone that contains brown to black, subhorizontal to horizontal, irregular, argillaceous laminae. Individual convex to concave, wispy, laminae are laterally continuous for 5 cm and give the rock a rubbly, brecciated, nodular appearance (Fig. 10B).

Unit B1 is comprised of 40 to 45% of < 4 microns wide, sucrosic dolomite, 20 to 25% white, 40 micron wide, euhedral dolomite rhombs, 15% anhedral micrite and up to 10% of a white pore filling anhedral dolomite cement. Greenish brown clays, concentrated along stylolites, form 15% of the rock. Ferroan dolomite and gypsum comprise up to 1 to 2%.

UNIT B2

Unit B2 has a gradational and highly irregular upper contact. This unit comprises a semirecessive to semiprominent, buff to orangey buff, thin to medium bedded, calcareous very finely crystalline dolostone that contains discontinuous wispy laminations, horizontal stylolites and white dolomite filled vugs (Fig. 10D). Horizontal to subhorizontal stylolites are faulted against vertical to subvertical fractures that are filled with dolomite. In section H, the rock is pale greenish grey and contains pale light grey clumps that

are comprised of 70% dolomite and 30% calcite. The matrix of the rock has a calcite to dolomite ratio of 50/50.

As a whole, the unit comprises 50%, 10 micron wide dolomite. Where the dolomite rhombs are in contact with one another anhedral crystals form; however, isolated crystals are euhedral. Highly corroded, subhedral rhombs of white dolomite (up to 550 microns wide) form a minor constituent. Bitumen and organics form 5% whereas other opaques, that are concentrated in the dolomitized areas, represent 10% of unit B2.

Unit B2 has undergone a considerable amount of dissolution. Lines of dissolution pervade throughout the unit, resulting in a "soft sediment" type of appearance (Fig. 10C). The unit has a wavy bedding character which is suggestive of stromatolites, however, no internal structure could be identified (Fig. 9E). In indistinct antiforms of section A chert replaces bedding surfaces in unit B (Fig. 10E).

UNIT B3

Unit B3 has a gradational upper contact that is representative of a disconformity (diastem). This unit comprises a semirecessive to semiprominent, light to medium grey, orangey buff, planar to slightly wavy, thinly bedded to medium bedded, calcareous, finely crystalline dolostone that contains medium to dark brown, highly argillaceous patches (Fig. 11A). These

FIGURE 11

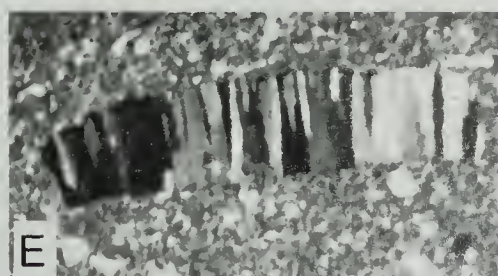
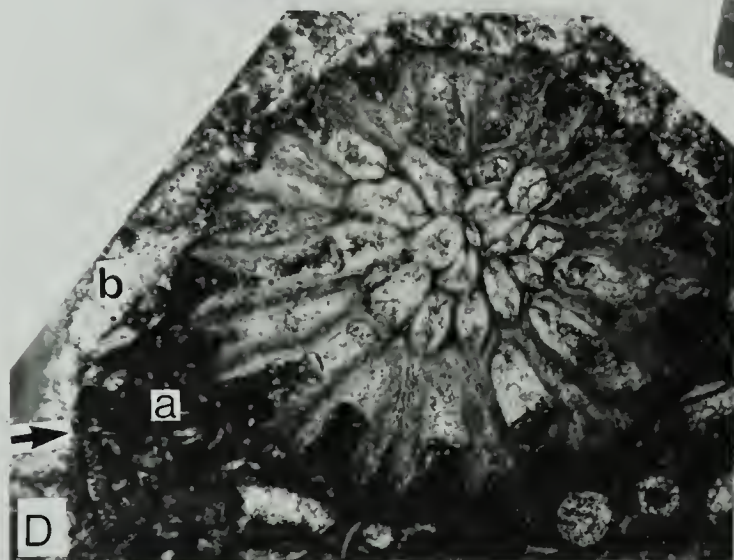
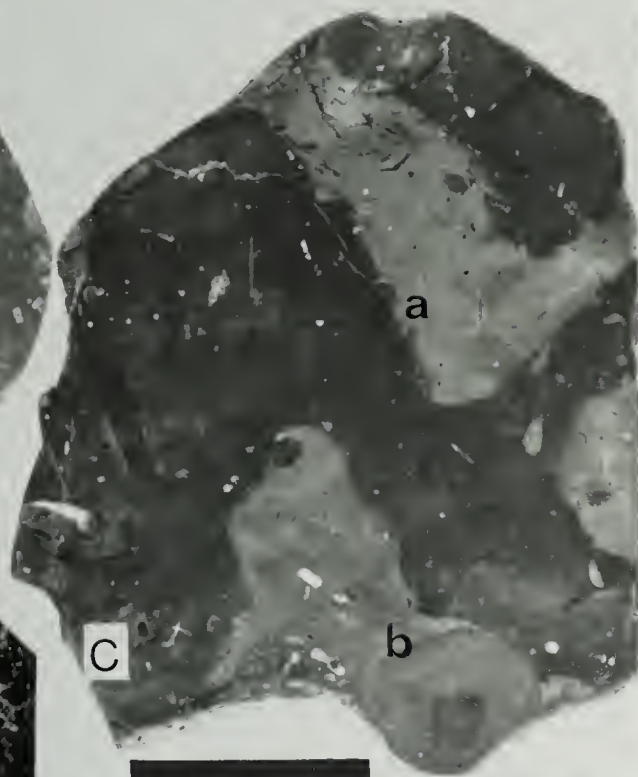
11A) Hand specimen of a calcareous, finely crystalline dolostone of unit B3 (section J) showing the argillaceous patches (darker areas). A gypsum mould that displays a prismatic head and swallow-tail end is filled with a white dolomite (a) that becomes clearer towards the center of the mould. Scale bar equals 3 cm.

11B) Hand specimen of an extremely bioturbated, very finely crystalline, fossiliferous, crinoidal, rubbly limestone from unit C1. The matrix (a) is darker with the patches (bioturbated areas) forming the lighter portions of the sample. Echinoderm fragments are abundant. Scale bar equals 3 cm.

11C) Hand specimen of a bioturbated, very finely crystalline, fossiliferous, crinoidal, rubbly limestone from unit C1. The patches (a, burrowed areas) are lighter than the matrix and contain dense micritic balls that have dark nuclei (b). Scale bar equals 3 cm.

11D) Photomicrograph of a lithoclast from unit C2 that is comprised of a colonial tabulate coral fragment and micrite (a). The matrix (b) of the rock sample has been aggraded to a 7 micrometer wide calcite. The arrow marks the edge of the lithoclast. Thin section under plane-polarized light. Scale bar equals 0.5 mm.

11E) Photomicrograph of a drusy cement lining a shell fragment (from unit C2). Thin section under cross-polarized light. Scale bar equals 0.5 mm.



patches have an indistinct enterolithic texture and are comprised of 50% 7 micron wide anhedral microspar, 50% 16 and 19 micron wide dolomite (two crystal modes), 1 to 2% ferroan dolomite and 3.5 mm diameter, spheroidal to elongate, well rounded vugs and fractures that are filled with a 310 micron wide dolomite. On a microscale, material appears to have been "plucked off" or ripped away from the patches, resulting in an irregular contact between the patches and the matrix. The fact that dolomite rhombs increase in size and number towards the edge of the patches, suggests that the patches are a diagenetic product. The matrix is formed of sucrosic euhedral 48 micron wide dolomite crystals and anhedral calcite crystals displaying a calcite/dolomite ratio of 1:3. Bivalve fragments are filled with fibrous calcite, and some shells are replaced with an isotropic mineral that is some form of salt. Spheroidal aggregates of chalcedony, that have a "felted" texture, are up to 225 microns in diameter. Swallow-tail gypsum moulds, lined with white sparry dolomite, were formed after two initial phases of dolomitization for they displace dolomite crystals in the matrix and the more argillaceous patches (Fig. 11A). Indistinct, subvertical microfaults also occurred after dolomitization because the dolomite crystals are terminated against the microfaults.

The rock is formed of 45%, aggraded 7 micron wide calcite, 15%, 16 and 19 micron wide dolomite, 30 to 35%, 45 to 50 micron wide dolomite and < 5%, 310 micron wide dolomite.

The upper contact of unit B3 is unconformable on a microscale. Bioclastic limestone of unit C is deposited on top of the previously dolomitized surface of unit B3. Oval microclasts, ripped away from the dolomitized unit B3 are incorporated into the base of the bioclastic limestone. Certainly, there must have been a significant time gap between the deposition and dedolomitization of unit B3 and deposition of the bioclastic limestone of unit C.

UNIT C

The biomicrite marker (below the Keg River marker) divides unit C into units C1 and C2.

UNIT C1

The upper contact of unit C1, with the biomicrite marker, is gradational over 10 to 15 cm. Unit C1 is a semiprominent to prominent, light grey to orangey buff to pale green, very thinly bedded to medium bedded, very finely crystalline, sparsely fossiliferous, crinoidal, rubbly limestone. Bedding is characterised by slightly wavy to planar contacts with beds thinning towards the top of the unit. The fresh surface color is light brown to medium brownish grey with dark brownish grey patches (Fig. 11B). The patches, representing 50 to 60% of the

unit, are generally elongate and irregular in form; however, towards the top of the unit they become circular (Fig. 11B). The patches represent lumps, comprised of microspar, embedded in a bioturbated, argillaceous microspar matrix. 5 cm thick argillaceous partings occur between the microspar lumps. Where bioturbation is extensive, the argillaceous material completely surrounds the patches. The darker color of the patches, along with the lighter color of the burrowed portions of the rock, give the unit a mottled/rubbly appearance. The burrowed areas, which are slightly dolomitized, contain a higher percentage of bioclastic fossiliferous debris and are generally more argillaceous than the patches. The burrowed areas are pyritized and are generally lined with red oxidized rims, with darker rims formed around the outer edge of the burrow. Iron stained calcite lined vugs, commonly occur throughout this unit.

The unit consists of 55% aggraded 8 micron wide calcite. White indistinct amorphous blebs of dolomite occur throughout the matrix. Towards the top of the unit coarse white dolomite replaces shell material. Stylolites, with maximum amplitudes reaching 127.0 microns, form 1 to 2% of the unit. Opaques and organic matter comprise approximately 4% of the rock.

Red rimmed peloidal micritic clasts constitute up to 2% of the rock. The lithoclasts, which have irregular

to rounded edges, are finer grained than the rest of the unit. In section B the long axes of the lithoclasts attain maximum lengths of 2.6 cm.

Well rounded micritic balls occur near the middle of the unit (Fig. 11C). The micrite forms a dense, featureless mass around a black colored micritic nucleus. Sharp, well defined contacts occur between the outer edges of the balls and the matrix. The balls may represent algal related structures or micritized burrow structures. If algae were present in the rock, the extensive recrystallization of the rock could easily account for the destruction of any algal features.

Unit C1 is fossiliferous, with the diversity and abundance of the fossils generally increasing towards the top of the unit. The upper part of the unit is characterized by numerous crinoid ossicles, abundant fossil hash, intact brachiopods and corals. Highly abraded and disarticulated pieces of echinoderm, bryozoans, conispiraled and planispiraled gastropods, trilobites, brachiopods, bivalves, ostracods, tentaculitids, conodonts, sponge spicules and solitary and colonial corals occur in varying proportions. The brachiopods are concentrated in the argillaceous partings between the microspar patches. The maximum fossil content is 30 to 40% of the rock. Crinoids generally form 10 to 15% of the rock, but towards the west (section 0) they become slightly rarer (7-10%). A

FIGURE 12

12A) Photomicrograph of a micrite lithoclast from unit C2 (white arrow marks the edge of the lithoclast) showing micritization (a) of a well preserved crinoid ossicle in the lithoclast. The aggraded calcite matrix of the rock sample contains fossil fragments with relict micrite rims (b). Thin section under plane-polarized light. Scale bar equals 0.5 mm.

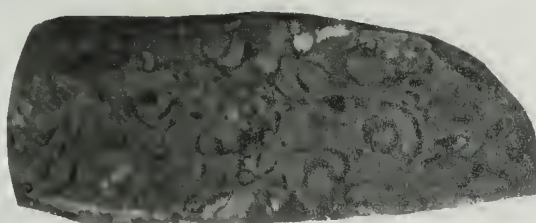
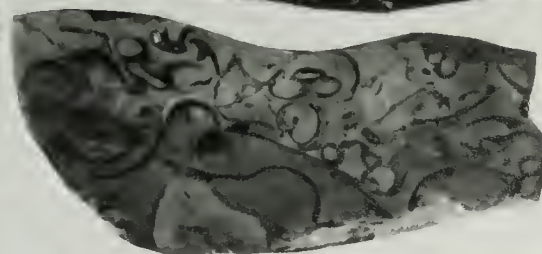
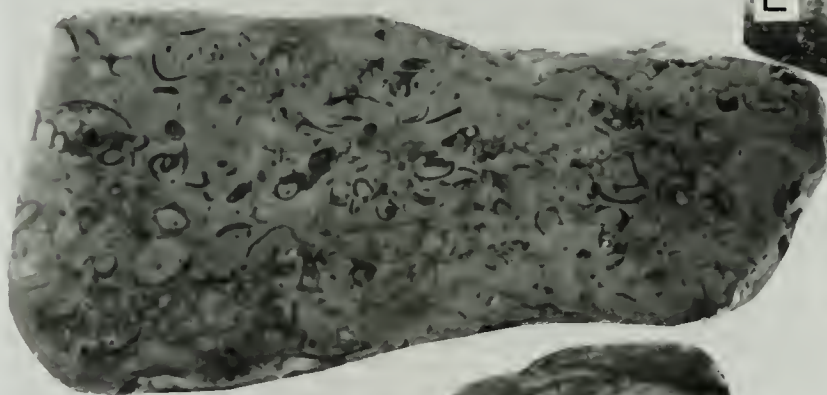
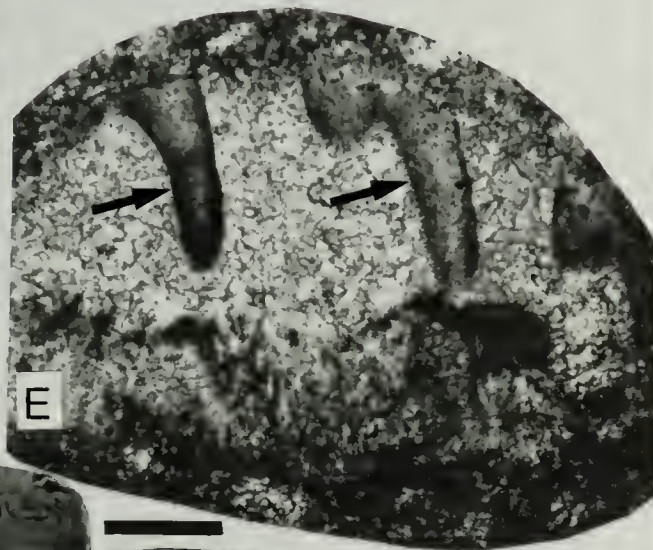
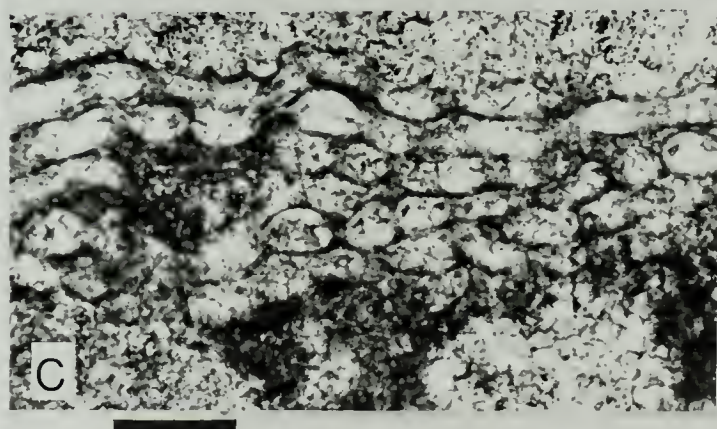
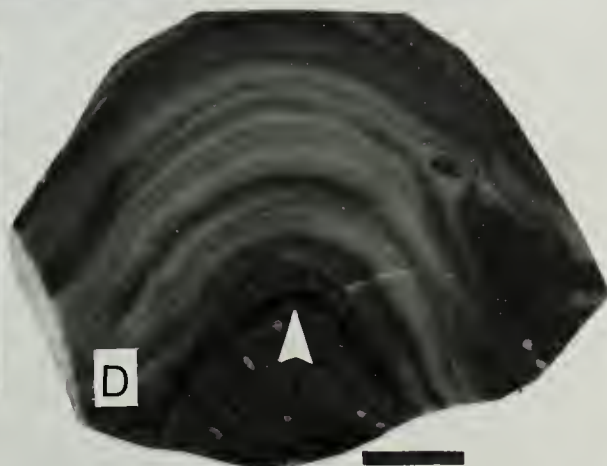
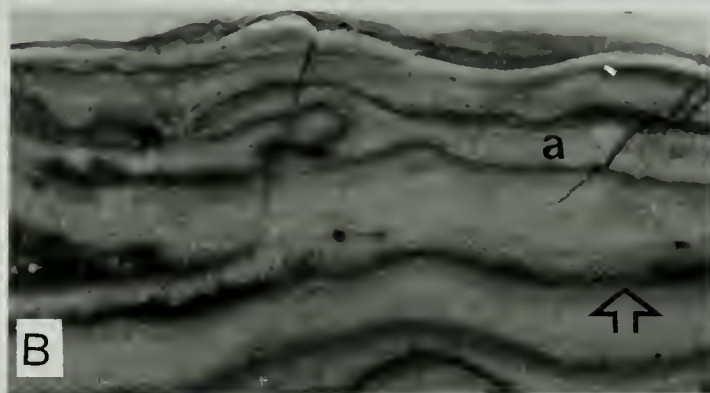
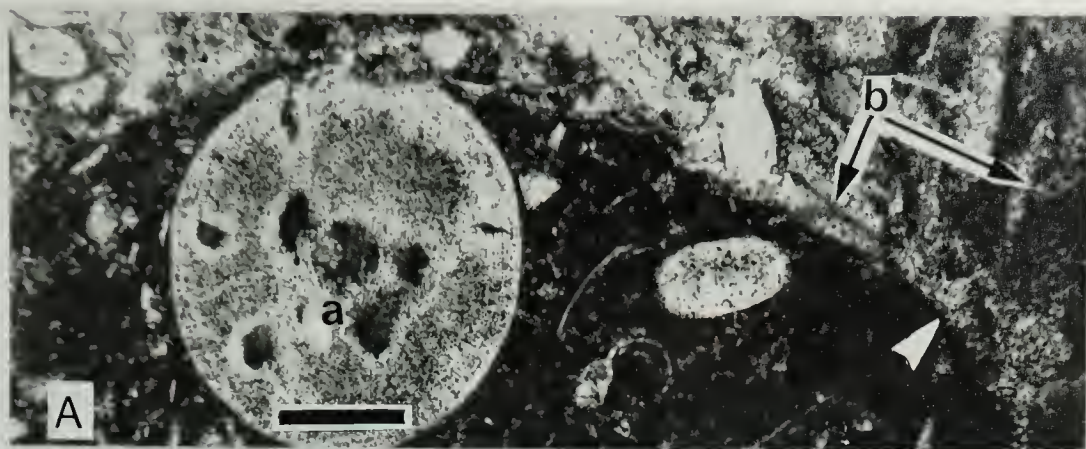
12B) Hand specimen of Alveolites (unit C) showing darker argillaceous laminae (arrow) which represent interrupted growth of the colonial coral. Sphaerocodium (algae) encrusts the coral (a). Scale bar equals 1 cm.

12C) Photomicrograph of Sphaericodium of 13B, depicting the chain-like texture of the algae. Thin section under plane-polarized light. Scale bar equals 0.05 mm.

12D) Hand specimen of a stromatoporoid encrusting an Atrypa shell (from unit C). Arrow is at lower edge of shell. Scale bar equals 1 cm.

12E) Photomicrograph of borings in fossil fragment. Arrows point to the borings. Thin section under plane-polarized light. Scale bar equals 0.05 mm.

12F) Hand specimens of the biomicrite marker. Samples taken from sections J (top left), F (lower left), A (top right) and O (lower right). Note the abundance of plani-spiraled and conical-spiraled gastropods. Scale bar equals 3 cm.



colonial coral zone which occurs near the top of unit C1, can be correlated throughout the exposed sections (Appendix 1).

Based on color differences and the abundance of burrows this lower unit can be divided into two units. The two units are easily recognized in sections F and G. However, the distinction is not as evident in the other sections.

UNIT C2

The contact between the biomicrite marker and the overlying unit C2 is gradational over 5 cm. Unit C2 is lithologically identical to unit C1 except that there is very little dolomitization of the burrows. The unit comprises a prominent, pale orange buff to medium grey, thinly bedded to medium bedded, aphanitic to very finely crystalline, fossiliferous, rubbly limestone. The fresh surface color is pale brown to medium and dark brown. Greenish hues occur in sections located towards the west end of the study area. Unit C2 typically displays planar to slightly wavy bedding. Bedding thickness varies from 1 to 30 cm throughout the unit, with a modal bedding thickness of approximately 5 cm. Although the unit is bioturbated, it is not as extensive as in unit C1. However, beds containing 50 to 60% burrows do occur near the top of the colonial coral zone in section J.

Lithoclasts, generally composed of an aphanocrystalline limestone, are more common in unit C2

than in unit C1 (Figs. 11D, 12A).

Unit C2 contains the same diversity of fossils as unit C1. However, there is a greater abundance of fauna in unit C2. The greatest diversity and abundance of fossils (especially corals and bryozoans) occur in the centrally located sections H and J. Echinoderms are common near the base and near the very top of unit C2.

Algal structures are clearly more evident in unit C2 than in unit C1. Coralline algae, commonly found encrusting Alveolites and Atrypa (Figs. 12B, 12C), are identical to the encrusting chain-like Sphaerocodium described by Machielse (1972). In the Alveolites colonies, periods of little or no coral growth are marked by Sphaerocodium and argillaceous layers (Fig. 12B). Interrupted growth of the Alveolites suggests that the environment was subjected to rapid changes, changes which could occur on a very shallow, moderately high energy carbonate platform. Machielse (1972) stated that Sphaerocodium occurs between the Lower Keg River Member and the basal beds of the Rainbow Reef Member. Machielse (1972, p. 218) added that the "Sphaerocodium may have played an important role in stabilizing the mounds upon which the Rainbow Reefs developed". This algal form has a modern equivalent that is found in similar environments (Machielse, 1972). The Recent form, in present day lagoonal and protected platform and/or bank environments, are common from low tide level to areas

with a water depth of approximately 11 m (Machielse, 1972). In unit C2, an alga displaying a cross-hatched algal fabric may be of a Dasycladaceae type. Another structure of probable algal origin occurs in unit C2 where borings in mollusc fragments may have been formed by algae (Fig. 12E).

BIOFACIES OF UNIT C

The lithological similarity of the units throughout unit C made correlation of the separate sections difficult, especially if the biomicrite marker was not apparent. However, biofacies of Atrypa, Alveolites and bivalves aided in the correlation of the sections along the Salt River (Appendix 1). Definite concentrations of the three fossil types occurred in most of the sections. In places, the fossils were highly concentrated and the biofacies could be easily identified (Appendix 1, section B). However, in other sections (Appendix 1, section K), the fossils were not as concentrated and delineation of the base and top of the biofacies was more difficult. Generally all the fossils occurred in their life or growth positions.

THE BIOMICRITE MARKER

Stratigraphic correlation of sections A through O was aided by a 0.25 to 0.3 m thick, fossiliferous, rubbly, recessive to semirecessive, light olive to pale greenish grey, packed biomicrite (Figs. 12F, Appendix 1). The lower and upper contacts are gradational with units C1 and C2, respectively. The unit is extremely rubbly with micritic

lumps (2 to 4 cm longest axis) that are separated by irregular argillaceous interlayers. Argillaceous content increases towards the northwest so that the maximum argillaceous content occurs in section O.

The rock comprises an anhedral mosaic of calcite with a modal crystal size < 4 microns. Ferroan calcite is common. Fossils are replaced with pyrite and traces of copper.

Sparry calcite, that displays irregular to straight intercrystalline boundaries, commonly occurs in the gastropod shells. The calcite crystal size increases towards the center of the shell possibly reflecting two to three episodes of cement generation. The gastropod shells are more susceptible to dissolution than the other fossil types because some shells (brachiopods and bivalves) still display a fibrous calcite texture, suggesting dissolution of the shells has not yet occurred. Discontinuous stylolites occur throughout the unit.

The fossils occur in a well lithified, approximately 5 cm thick band in the rubbly biomicrite marker. This fossiliferous band occurs at 4 localities along the Salt River (sections A, B, J, and O; Fig. 12F). Although the marker horizon occurs in sections E, F and G the fossiliferous bands do not. Where the marker does contain fossils, they include: rare trilobites, tentaculitids, bryozoans, plani-spiral gastropods, star, circle and two-holed (Gasterocoma) crinoid ossicles, ostracods and algae. Algae are represented by 3 mm diameter ovate

structures, by coralline algae, and by thick, dense, indistinctly layered micritic rims formed around the allochems. Thin shelled brachiopods and conical gastropods generally form 5 to 30% (each) of the fossiliferous band. Conical gastropods are concentrated in all sections containing the fossiliferous bands, but brachiopod abundance increases to the east. The fossiliferous band contains 30 to 65% fossils. The fossils are generally slightly abraded, however articulated forms are also common.

IV. DIAGENESIS AND DEPOSITIONAL HISTORY OF THE SALT RIVER STRATA

A. DIAGENESIS

UNITS A AND B

Unit A originated as an argillaceous micrite. The original lithology of unit B is not certain, but ghosts of allochems in the unit and remnant calcite inclusions in the dolomite suggest that it was once a limestone. The original lithology of unit A is more discernible than it is in unit B because the latter unit is diagenetically more altered. Dissolution of shells is evident in unit A. Subsequent pore space was filled with an aragonite or high magnesium calcite cement that increases in size to the center of the void. Some shells still retain the fibrous nature of the original shell structure thus suggesting that they were not dissolved. Although the rock of unit A retains a modal crystal size < 4 microns, aggradation of the micrite was extensive and resulted in aggraded micrite envelopes around the allochems and a peloidal texture (Fig. 13A). The peloids are remnants of the original lithology, whereas the calcite surrounding the peloids is a product of neomorphism. The peloids may represent faecal pellets; however, the variable size, the lack of an ill sorted agglomeration of various kinds of fine particles and the lack of organic matter (some of the criterion used for defining faecal pellets from Bathurst, 1975) suggest that these spherical structures are

not true pellets.

DOLOMITIZATION

Units A and B underwent 3 phases of dolomitization. However, dolomitization is more extensive and more distinct in unit B.

Phase 1 dolomitization, which represents a replacement of the original micrite, produced a dirty, inclusion rich, anhedral to subhedral, sucrosic dolomite. From the base of unit A to the top of unit B the phase 1 dolomite increases in size. At the base of unit A the dolomite has a width of 2.5 microns, which increases to slightly < 4 microns in unit B1, to 10 microns in unit B2, to bimodal dolomite sizes of 16 and 19 microns in unit B3 (Fig. 14B).

Phase 1 dolomite is partially replaced by a coarser, clearer dolomite that is the result of phase 2 dolomitization. Size gradation of phase 2 dolomite is not as evident as it was for phase 1 dolomite, however, there is an indistinct overall increase in phase 2 dolomite from 44 microns wide in unit A to 48 microns wide at the top of unit B3 (Fig. 14B). Phase 2 dolomite increases in abundance from $< 10\%$ in unit A to 36% in unit B. The dolomite generally forms as subhedral crystals in a sucrosic matrix, however, where individual dolomite rhombs are not in contact with other rhombs, and no interference of crystal growth occurs, the resultant crystal is perfectly euhedral.

Phase 3 dolomitization produced dolomite crystals up to 300 microns wide. This dolomite is clear, white, sparry,

FIGURE 13

13A) Photomicrograph of sample from unit A showing micrite, aggraded coarser calcite (dark areas) and dolomite (light areas). Grumeleuse blotches of micrite (a) give unit A a peloidal texture. Euhedral dolomite is replaced by calcite (b). Thin section is stained and under plane-polarized light. Scale bar equals 0.1 mm.

13B) Photomicrograph of sample from unit B3. Darker areas are calcite, lighter areas are dolomite. Some dolomite rhombohedrals are completely replaced with calcite (a) whereas other rhombs are only partially replaced (b). Large single crystals of calcite replace the previously formed dolomite but the edge of the euhedral dolomite rhomb remains (c). Dissolution began near the base of the large euhedral dedolomite rhomb (d). Thin section is stained and under plane-polarized light. Scale bar equals 0.1 mm.

13C) Photomicrograph of sample from unit B1. Euhedral rhombs of dolomite are partially to completely replaced by calcite. Thin section stained and under plane-polarized light. Scale bar equals 0.1 mm.

13D) Photomicrograph of sample from unit B3 showing calcite (dark area) replacement of dolomite (light area) along a dedolomitizing front, where the direction of the advancing front is shown by the straight arrows. Calcite replacement is irregular along the front (a). A uniform mosaic of calcite crystals has replaced the dolomite rhomb in the center (b) with euhedral dedolomite occurring throughout the calcite pocket. The darker, deeper red of the calcite suggests a higher concentration of calcium along the dedolomitizing front (c). Thin section stained and under plane-polarized light. Scale bar equals 0.1 mm.

13E) Photomicrograph of rock from unit B2. Dolomite rhomb with cleavage (a) was partially replaced with calcite (b), but some original dolomite remains (c). Dissolution of the central portion of the dolomite has occurred, thus creating pore space (d). Thin section stained and under plane-polarized light. Scale bar equals 0.1 mm.

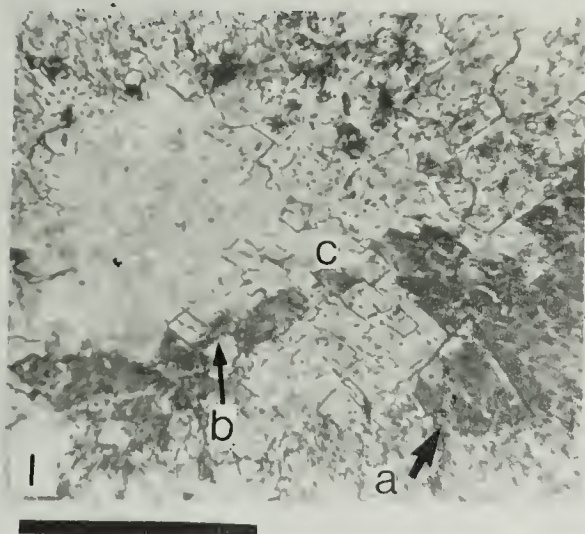
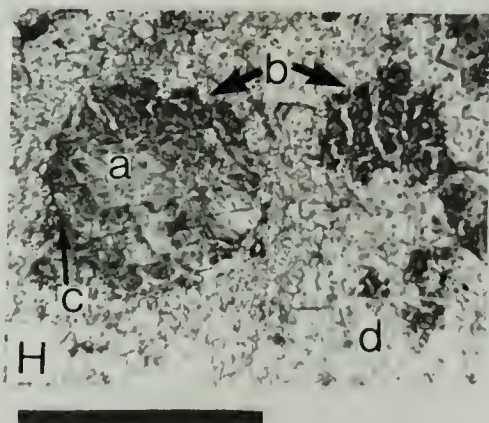
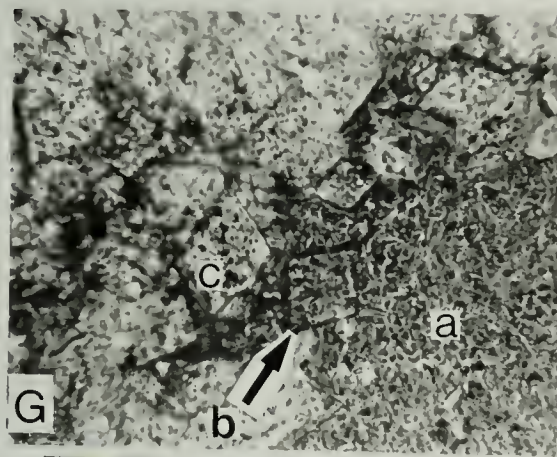
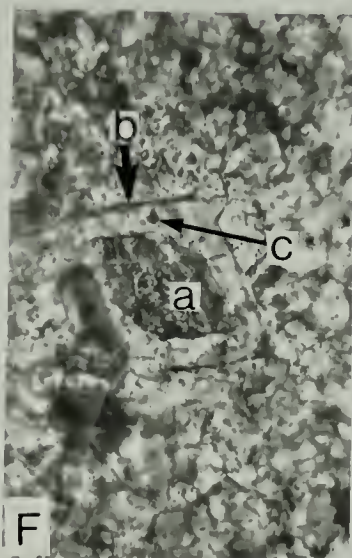
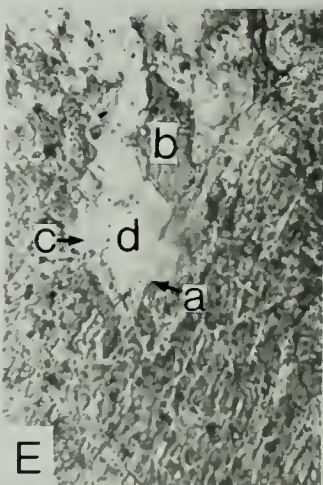
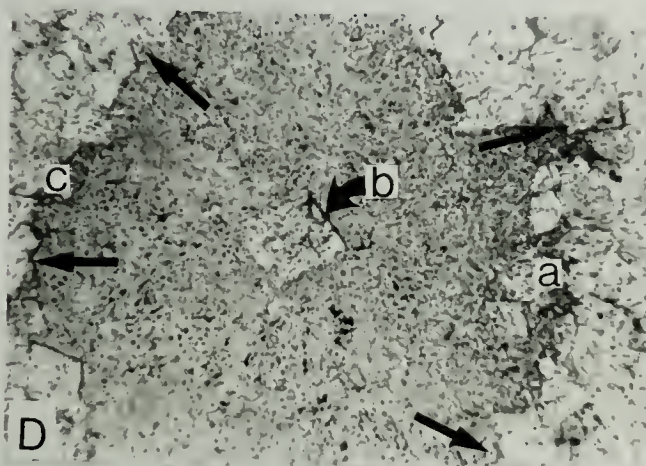
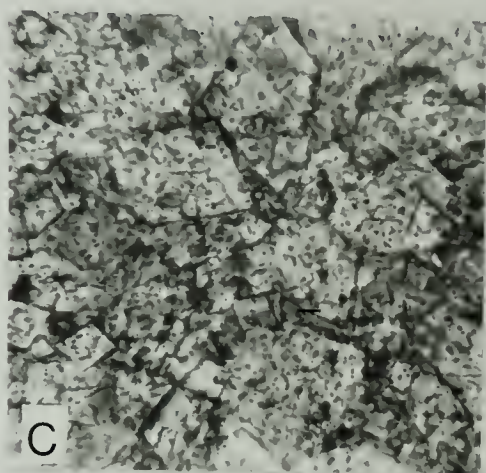
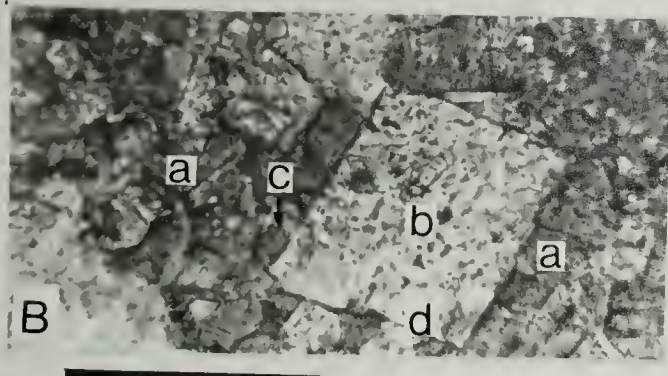
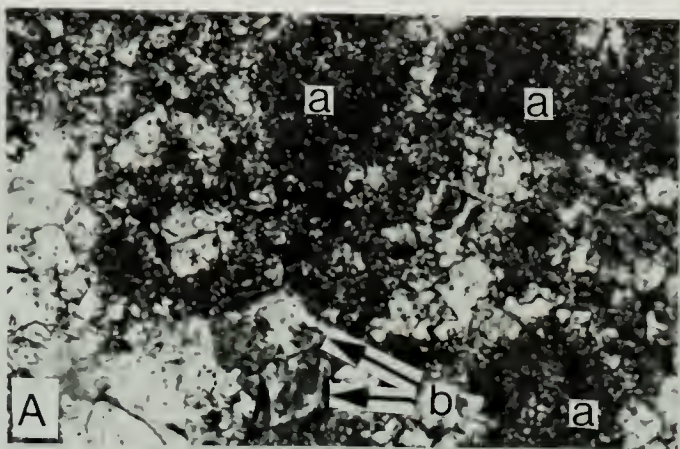
13F) Photomicrograph of sample from unit B2 showing a "zoned dedolomite" (a). This dedolomite represents the initial phase of centripetal replacement of dolomite (b) by calcite. Replacement occurred along cleavage planes in the dolomite with nucleation of the replacing calcite probably having occurred on relict calcite (c) in the dolomite. Thin section is stained and under plane-polarized light. Scale bar equals 0.1 mm.

13G) Photomicrograph of sample from unit B3. Calcite (a) replaces dolomite (light areas)

with a anhedral mosaic texture developed. Some dolomites have been completely replaced (b) whereas others are only partially replaced (c). Thin section stained and under plane-polarized light. Scale bar equals 0.1 mm.

13H) Photomicrograph of a sample from unit B2 showing calcite (dark areas) having replaced dolomite (light areas). Calcite replaced the dolomite rhombs (a) having formed round peloids (b). This dedolomitizing process produced a grumeleuse texture. Calcite tends to have replaced the dolomite along the edges and along cleavage in the dolomite, however, rhomb boundaries can still be identified (c). Dedolomitization had started to form another peloid structure in the lower corner (d). Thin section is stained and under plane-polarized light. Scale bar equals 0.1 mm.

13I) Photomicrograph of sample from unit B1 showing end members of the dedolomitization process. Dedolomitization is a gradational process where dolomite rhombs can be either replaced completely by an anhedral mosaic of calcite (a) or only partially replaced by calcite along cleavage planes (b). Dissolution occurs near the top of the larger rhomb. Thin section stained and under plane-polarized light. Scale bar equals 0.1 mm.



Diagenesis Prior to Deposition of Unit C

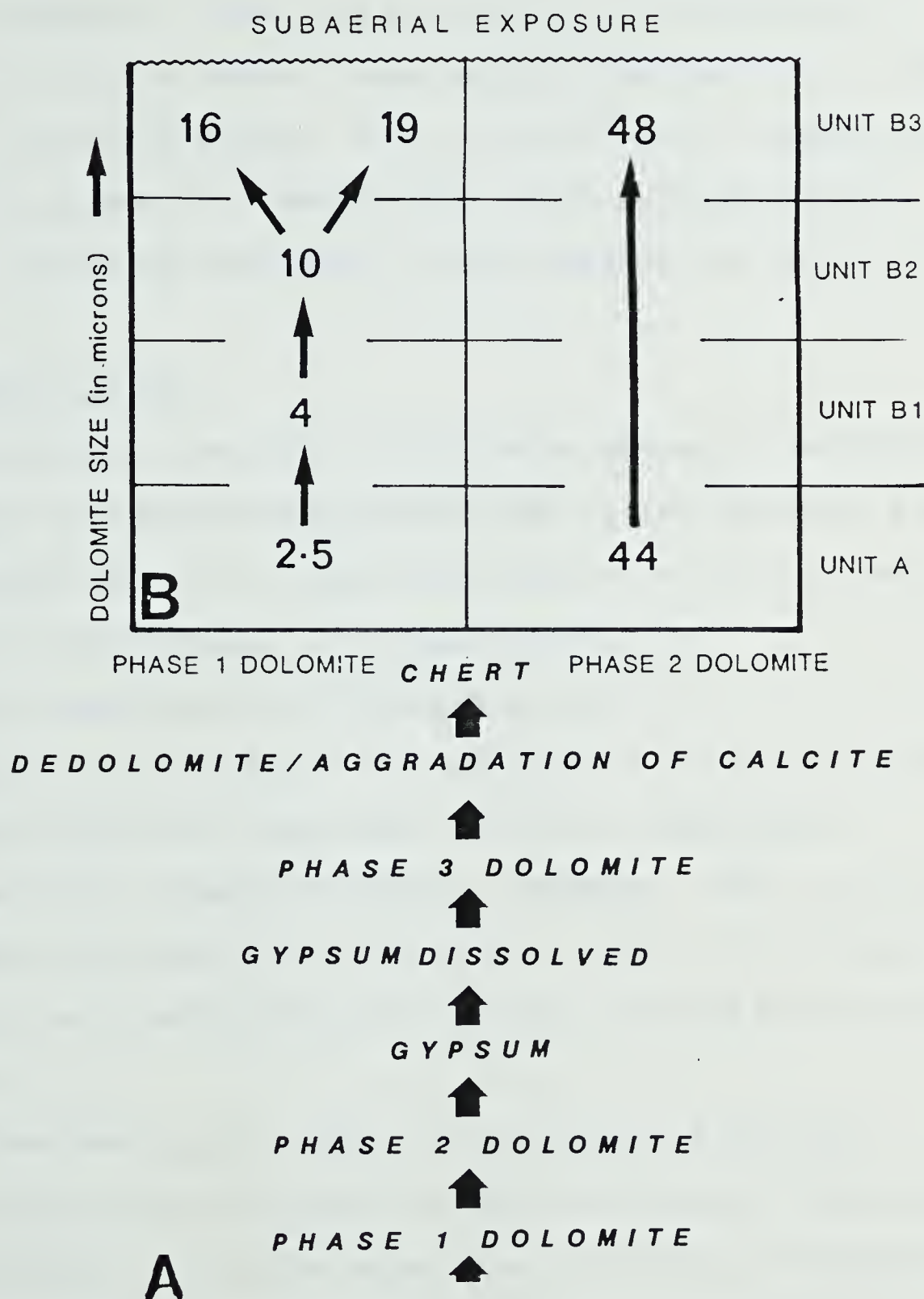


FIGURE 14. Diagenetic Processes Related to Subaerial Exposure following Deposition of unit B3.

displays planar intercrystalline boundaries, a "brushed texture" and fills previously formed gypsum moulds in unit B. The gypsum is indicated by swallow tail moulds and prismatic crystal ends. These moulds displace phase 1 and phase 2 dolomite along 1 to 2 cm microfaults, suggesting that the gypsum was formed after these dolomite phases. Phase 3 dolomite increases in size towards the center of the mould.

DEDOLOMITIZATION

Extensive corrosion of all three phases of dolomite followed the dolomitizing phases that effected units A and B. Aggradation of the remaining calcite in units A and B occurred simultaneous with dedolomitization.

The replacement of dolomite by calcite (dedolomitization) was first reported by Von Morlot (1848). The term dedolomite describes partly or completely dedolomitized rhombs of calcite (Goldberg, 1967), or can be applied to metamorphic rocks (Teall, 1903). This thesis will use the term dedolomite as originally defined by Von Morlot (1848).

Shearman et al, (1961, p. 2) suggested that the following criteria be met in order to classify limestone as dedolomites. "i) By the occurrence of relics of incompletely replaced dolomite crystals. ii) By the presence of pseudomorphs of calcite after dolomite. iii) By the existence of palimpsest textures in which rhombic zones of ferric oxides or grain boundaries of the earlier dolomite

crystals remain as ghosts within the new generation of calcite".

In the Salt River sections all 3 criteria are met by rocks in units A and B. Both partial and complete replacement of the dolomite is obvious. Dolomite rhombs displaying good cleavage are only partially replaced by calcite; elsewhere complete replacement occurs, with only the rhombohedral shape of the dolomite left as proof of the pre-existing dolomite (Fig. 13). A series exists where normal dolomite forms one end member and completely calcite replaced dolomite rhombs form the other end member. The third criterion, proposed by Shearman et al. (1961), exists but is not as evident in the Salt River rocks as the other two criteria. There are no rhombic zones of ferric oxides, however, partly replaced dolomite rhombs occur in large masses of calcite. The edge of the dolomite rhomb can be traced through the calcite, thus forming a ghost of the crystal edge even though the calcite has replaced dolomite on either side of the rhomb boundary (Fig. 13).

In unit B, dark, rhombic, calcite centers in the dolomite rhombs give the dolomite a "zoned appearance" (Fig. 13F). Based on Evamy's (1967) scheme of dedolomitization, the outer rim of dolomite could later be replaced with a coarser, inclusion free, slightly cleaner calcite, thus resulting in a grumeleuse texture. Shearman et al. (1961) and Evamy (1967) reported clotted or grumeleuse textures from dedolomitized rocks suggesting that where dolomite is

replaced by calcite a grumeleuse texture results. The dedolomites are formed of rhombohedrons with dark centers, rich in calcite inclusions and surrounded by clear, inclusion-free calcite rims. Shearman et al. (1961) suggested that the grumeleuse texture is due to aggradation of dolomite into a system of clots and stringers. A grumeleuse texture occurs in both calcite and dolomite in the Salt River strata. In units A and B, the texture is due to aggradation of a micrite or a dolomicrite into a coarser calcite or dolomite, respectively.

Rhombohedral pores are common in dedolomitized rocks and Evamy (1967) gave good evidence to show that the rhombohedral pores are developed through the selective leaching of dedolomite and not by leaching of dolomite. Rhombohedral pores are numerous in unit B and represent leaching of dedolomites, not dolomite. Evidence supporting this statement is given where well developed euhedral rhombs are filled with a mosaic of equidimensional calcite where only partial dissolution of the calcite has occurred. A critical observation by Evamy (1967), from a similar situation, is that if this were a calcite cement filling a pre-dissolved dolomite or dedolomite rhomb, then the crystal size of the remaining cement in the pore would increase towards the center of the pore. The origin of the previous rhomb could not be identified. However, where this does not occur and the calcite is equidimensional it is unquestionably a product of dedolomitization. Such is the

case for the rhombic calcite from units A and B. The partially dissolved rhombic pores of units A and B display no evidence of calcite filling cements, but do display anhedral equidimensional mosaics of calcite; thus suggesting that the pores represent dissolution of dedolomite (Fig. 13). Dedolomite calcite is leached before other calcite in the rock because it is generally comprised of high magnesium calcite or aragonite and is therefore mineralogically unstable (Evamy, 1967).

Three types of dedolomitization textures were reported by Shearman et al. (1961, p. 3), "1) Those rocks in which the individual crystals of dolomite have been replaced by a finer grained mosaic of calcite crystals. 2) Those rocks in which the dolomite has been replaced by a mosaic of calcite crystals which are coarser grained than the earlier dolomite. 3) Those rocks in which the calcite replacement mosaic has tended to regenerate the pre-dolomitization texture of the limestone."

The first texture is the most evident type of dedolomite replacement in the Salt River strata. Well defined rhombs of dolomite have been replaced with a fine grained equidimensional mosaic of calcite (Fig. 13). Shearman (1961) and Evamy (1967) termed such calcite-replaced rhombohedrals "composite calcite rhombs", further stating that they are pseudomorphs of calcite after dolomite. Such rhombs are common in units A and B of the Salt River strata. The growth of calcite in dedolomite

textures generally begins with nucleation of calcite on calcite inclusions derived from the original limestone and carried over through the intermediate dolomite stage (Evamy, 1967). Relict calcite inclusions in the replacement dolomite may have provided nucleation sites for the dedolomites in the Salt River strata.

Shearman et al. (1961), in agreement with Khvorova (1958), stated that the replacement process can be either centrifugal or centripetal. Centrifugal replacement produces dolomite rhombs scattered through a limestone groundmass, whereas centripetal replacement is where larger calcite crystals poikilitically enclose the dolomite. Both types of replacement occur in units A and B, however centripetal replacement is less common. In the Salt River strata, however, calcite does replace the outer periphery of the dolomite rhombs, forming embayments and highly corroded crystal edges. The calcite from the Salt River strata has a corroding, as well as a replacing effect (Fig. 14B).

Dedolomitization is indicative of subaerial exposure (Shearman et al., 1961; Evamy, 1967; Goldberg, 1967; De Groot, 1967). De Groot (1967) showed that dedolomitization can only occur at or near the earth's surface where the carbon dioxide content and degree of temperature are relatively low. He suggested that dedolomitization is ineffective at temperatures higher than 50° C and PCO greater than 0.5 atmospheres. High Ca/Mg ratios and high rates of water flow, in association with low PCO and low

temperature, are needed to produce dedolomite (De Groot, 1967). Folk and Land (1975) also suggested that dedolomite forms under high Ca/Mg and low salinity conditions. The above restrictions suggest that dedolomitization is not a subsurface phenomena but solely the result of near surface exposure.

Von Morlot (1848) originally suggested that dedolomitization is due to the interaction of gypsum and dolomite where sulphate ions are supplied by the leaching of gypsum. Tatarsky (1949) rejected Von Morlot's views and stated that dedolomitization was hampered by the presence of sulphate. De Groot (1967) stated that when PCO was low gypsum could play an active role in dedolomitization. There is no gypsum in unit B3, however, swallow tail gypsum moulds and indistinct enterolithic folds in the dolostone represent relict gypsum textures. It is possible that the gypsum did aid in the dedolomitization process in the Salt River strata.

Units A and B of the Salt River area undoubtedly met the three criteria defining dedolomites, as defined by Shearman et al. (1961). Textures such as rhombohedral pores and grumelleuse structures, typically associated with dedolomites, also occur in units A and B. Therefore, the above evidence clearly suggests that a major proportion of the calcite in units A and B is dedolomite. Based on previous literature, and in association with additional evidence of subaerial exposure, this dedolomite is thought

to have been produced during an interval of subaerial exposure of units A and B during the Middle Devonian.

UNIT B DIASTEM

The following evidence supports the existence of a diastem at the top of unit B3 and subaerial exposure of unit B during the Devonian.

The extensive dolomitization and dedolomitization phases that altered units A and B did not occur in unit C because unit C was deposited after the dolomitizing and dedolomitizing phases (Figs. 14A, 14B). The basal few centimeters of unit C contain rip up clasts of the dolomitized and dedolomitized rock of unit B (on a microscale). The above evidence suggests that there is a period of non-deposition between the dolomitizing and dedolomitizing events of unit B and the transgressive event depositing the limestone of unit C. The minimal amount of time reflected by this interval of non-deposition was the time needed to completely dolomitize and dedolomitize units A and B prior to the transgressive event that deposited unit C. The diastem at the top of unit B represents a subaerial exposure surface.

Throughout unit B evidence of subaerial exposure is given by, the dissolved appearance of unit B2, the chert replacement of unit B2 (section A), the dolomitization and subsequent dedolomitization, the varying thickness of the units (eg. the pinch-out of unit B1) and the presence of wavy bedding. The wavy bedding of unit B2 appears

stromatolitic, however, it does not represent algal formed stromatolites produced by sediment trapping, binding or precipitation as a result of growth and metabolic activity of microorganisms. The wavy bedding is due to dissolution at a subaerially, or nearly subaerially, exposed surface. Read (1976) suggested that some abiogenic stromatolitic structures are products of weathering associated with an unconformity surface. The above textures represent subaerial conditions formed in response to a diastem that exists between units B and C.

UNIT C

Unit C contains fibrous rimming cements as well as triple junction contacts. These textures are evidence of cementation in a subtidal environment/active marine phreatic zone (Longman, 1980). There is very little dolomite and no dedolomite in unit C. The dolomite occurs only in the bioturbated portions of the rock and is not related to the dolomitizing events effecting units A and B. Unit C is less recrystallized and a higher percentage of fossils retain their fibrous nature compared to units A and B. The higher percentage of algae in the upper few meters of the section also suggests that extensive aggradation of the limestone did not occur.

A drusy cement which resembles Dunham's (1969) "massive dripstone" cement is common (Fig. 11E). Dunham (1969) suggested such a cement is typical of subaerial origin. In unit C, this cement is possibly the result of a later

diagenetic event that was related to another subaerial surface that occurred above the biomicrite marker of unit C. Such a surface may be related to dissolution and subsequent brecciation events which later caused deformation of the Keg River Formation along the Salt River.

B. A SEDIMENTOLOGICAL/DIAGENETIC HISTORY OF THE KEG RIVER FORMATION ALONG THE SALT RIVER

The Keg River Formation of the Salt River, Wood Buffalo National Park, represents an overall transgressive sequence, with minor regressive phases represented by the unit B diastem and the biomicrite marker (Fig. 15). Deposition of the Salt River strata occurred on the edge of the La Crete Basin in a very shallow carbonate shelf environment where there was a high degree of argillaceous input into the environment. The diagenetic events and depositional setting are similar to those described by Dunham (1969), Ferguson et al. (1982) and Warren (1982). All three authors suggested that subaerial exposure produced a fresh water lens below a previously lithified sediment. However, Ferguson et al. (1982) and Warren (1982) stated that the fresh water lens occurs beneath early cemented, expanded and cracked, lithified crusts of environments that contain pseudoanticlines. These environments more closely resemble the environment represented by the Salt River strata.

Unit A represents deposition in a reducing, subtidal environment. Organics were not oxidized and occur in varying

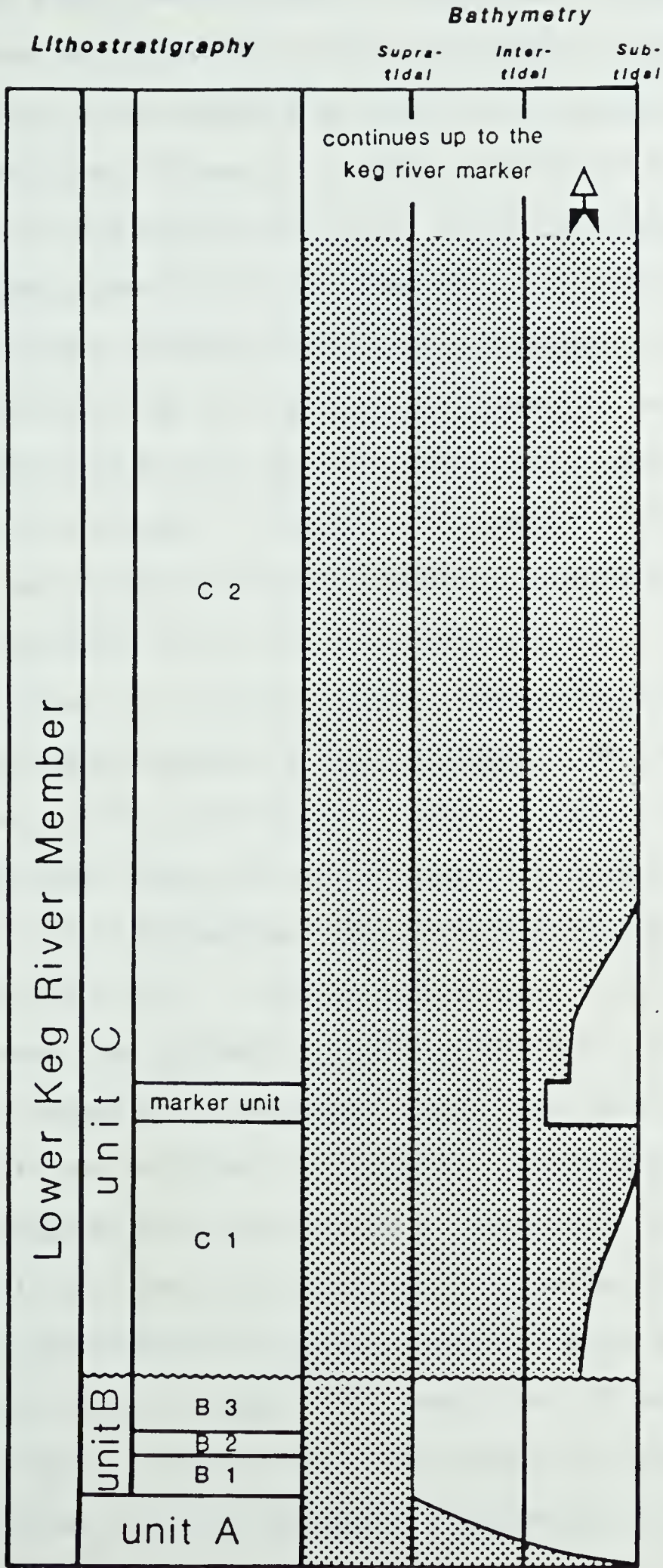


FIGURE 15. Suggested Bathymetry for the Salt River Lithostratigraphic Units.

percentages along laminations in the rock. A regressive sequence was initiated when the environment changed to an intertidal one, represented by unit B1. The undulatory, wispy, organic/argillaceous laminae of unit B1 reflects deposition in the intertidal zone. With continued regression unit B became subaerially exposed in a supratidal setting, thus forming the diastem that exists between units B and C. Prior to, and during this period of subaerial exposure the sediment was extensively dolomitized in an intertidal to supratidal environment. Swallow tail gypsum growth displaces the first two phases of dolomitization suggesting a brief period of hypersalinity near the top of unit B. The dolomite may have formed in a "mixing zone" (schizohaline environment) where meteoric water began to mix with hypersaline water. Schizohaline refers to an environment altering between hypersaline and near-fresh water conditions as is found in a floodable sabkha or phreatic mixing zone (Folk and Land, 1975). Further addition of fresh water occurred when the sediment migrated through time to the shallower, subaerially exposed, supratidal environment; thus resulting in an increase in dolomite size and clarity of the dolomite towards the top of unit B. Folk and Land (1975) suggested that a vertical change of dolomite represents a changing concentration of Mg/Ca and that such a dolomite probably forms as a result of a reduction of salinity. The coarser, clearer, dolomite of the latter two phases of dolomitization (especially phase 3 dolomite) is evidence of

a more fresh water environment. The increasing abundance and increase in size of phase 1 dolomite to the top of unit B may also represent a dolomite formed under the increasing influence of fresh water (Fig. 14B). "We conclude that dolomite can form in surface fresh waters with Mg/Ca ratios as low as 1:1 providing crystallization is slow enough for ordering to take place" (Folk and Land, 1975, p. 63). When hypersaline water is diluted with fresh water schizohaline conditions dolomite can form with Mg/Ca ratios as low as 1:1 (Folk and Siedlecka, 1974). The above suggests that dolomite can easily form under the increasing input of fresh water.

Dolomitization of unit B was followed by a period of diagenesis in a fresh water environment. With continual regression of the Devonian Seas, the domed carbonate strata exposed along the Salt River underwent a major phase of neomorphism where minor amounts of calcite in unit B were aggraded to a 5 to 10 micron diameter microspar. Land et al. (1967) introduced the term "stabilization" for a similar process where progressive calcitization of aragonite and high magnesium calcite occurred in the subaerial environment. Dunham (1969) suggested that "stabilization" in the Townsend mound lasted 100,000 years.

In unit B, diagenesis in a fresh water environment is also reflected by extensive dedolomitization. Dedolomitization is reflected by re-entry of dolomite rhombs (centripedal replacement) and centrifugal replacement of the dolomite rhombs by equidimensional mosaics of anhedral

calcite. This dedolomitization phase probably occurred synchronously with the aggradation of the calcite because both events would require fluids with low salinities and low Mg/Ca ratios (Folk and Land, 1975, Fig. 2, p. 62).

Dissolution of unit B, based on previously mentioned textures, also occurred during this period of subaerial exposure.

Following the period of subaerial exposure, a transgression of the Devonian seas ended the period of nondeposition. Rubbly limestone, of unit C, was deposited on top of the subaerially exposed unit B. The high amount of bioturbation and "micritized balls" suggest deposition in a subtidal, possibly intertidal environment. Dolomitization of the burrows suggests that the area reverted back to a marine influenced area with minor fresh water/marine water mixing occurring as the Devonian seas overlapped the Keg River strata. "Mixing zone" conditions were "short lived" and were evidently not as favorable for complete dolomitization of the rock as in unit B. However, dolomitization in this case could have simply been due to dolomitization by subsurface fluids at a later point in time. The intertidal/subtidal environment of unit C1 deepens towards the top of the unit for a colonial coral biofacies occurs just below the biomicrite marker.

The biomicrite marker represents a return to a regressive phase of the Devonian seas reflecting a restricted environment where conical gastropods

dominated. It may reflect a lagoonal facies that occurred landward of the supratidal flat to shallow subtidal conditions of the previously deposited units.

Unit C2 marks a deepening of the area during the Middle Devonian. A transgressive sea caused the lagoonal deposits of the biomicrite marker to be overlain by a highly fossiliferous rubbly limestone. The abundant fauna, the diversity of fauna and the rubbly nature of the rock suggests that reefoid to good mound forming conditions were being initiated at this time on the edge of the La Crete Basin. Corrigan (1975) stated that in the center of the La Crete Basin mounds were not initiated until the first stage of the upper platform unit (Upper Keg River Member, immediately above the Keg river marker). Deepening of unit C2 is also evident from the marine biofacies. Unit C2 represents a continually deepening environment that eventually deepened to the environment represented by the Keg River marker.

Although the Keg River marker has not yet stratigraphically been reached in the Salt River exposures, it is not far above the top of the exposed sections for it occurs in breccias exposed along the river and in unit C elsewhere in the park. This marker represents argillaceous, deeper water conditions that prevailed in an anoxic environment.

Subaerial exposure of carbonate mounds reported by Corrigan (1975) in the upper platform unit (Upper Keg River

Member) may be analogous to unit C of the Salt River strata. Subaerial exposure may also have occurred above the mound forming sections of unit C, along the Salt River (Chapter 6).

SYNOPSIS

The depositional/diagenetic history of units A through C, apart from the pseudoanticline structures, is as follows.

1. Sediment of unit A was deposited in a subtidal environment.
2. Micrite envelopes developed around the allochems.
3. Partial dissolution of the allochems occurred, with an aragonite, possibly high magnesium calcite cement filling the voids.
4. Regression occurred, with deposition of units B1 through B2 representing deposition in an intertidal to supratidal environment. Regression continued until unit B was subaerially exposed during the diastem developed prior to deposition of unit C.
5. Dolomitization of units A and B in a "mixing" zone environment. Continual regression and decrease in Mg/Ca ratio, due to increases of fresh water, caused a gradational change of dolomite up through units A and B. Dolomitization is represented by three phases:
 - a. Phase 1 dolomite - Small dolomite rhomb stage.
Dolomite is brown and dirty.
 - b. Phase 2 dolomite - Larger subhedral to euhedral white dolomite.

- c. Gypsum precipitation and subsequent dissolution.
 - d. Phase 3 dolomite - Sparry, clear, dolomite cement.
6. Continued regression and extensive dedolomitization.
Dedolomitization was synchronous with aggradation of the original calcite.
 7. Silicification - Chert replaced previously formed dolomite and dedolomite. Extensive dissolution of unit B occurs during subaerial exposure, but relative time cannot be determined.
 8. Fracturing - Fracturing events have occurred throughout the depositional history of units A and B. However, this stage is evident because the fractures are filled with a non-aggraded calcite cement that display straight planar intercrystalline boundaries.
 9. Transgression - Unit C1 deposited on top of highly diagenetically altered unit B.
 10. Biomicrite marker deposited. Marker represents a minor regression.
 11. Unit C2 deposited. Represents continued deepening up to the Keg River marker.

V. ANTIFORM STRUCTURES OF THE KEG RIVER FORMATION, SALT RIVER AREA

A. MORPHOLOGY

Gently folded structures occur along the Salt River in the slightly brecciated to extensively brecciated sections A, E, F, G and H (Fig. 16). Three sizes of antiforms occur. The smallest antiforms, in unit A (section H), are up to 0.15 m high and 0.5 m wide. The moderately sized antiforms PA1, PA2, PA3 and PA5 (PA refers to pseudoanticline, which is defined as "any antiform structure formed by the expansion of indurated beds where the process and environment responsible for expansion cannot be established" [Assereto and Kendall, 1977]) are less than 1 m high and 1.5 m wide (Figs. 16, Appendix 3). They are extensively brecciated antiforms that generally have faulted axial crests with fault planes dipping at approximately 70° and axial trends of approximately 156° . In sections E, F and G the larger antiforms (PA4, PA6, PA7 and PA8) are up to 3.6 m high and 5.4 m wide. The width to height ratio ranges from 1.5:1 to 2.4:1 with a mean width to height ratio of 1.9:1. The flank beds of the pseudoanticline structure dip at less than 10° near the base and 85° near the crest. Typically, the crest forms a pointed tepee-like shape (Figs. 16G, 17G), but some of the larger pseudoanticlines have rounded axial crests (Figs. 18, 19C).

Saucer shaped areas occur between the antiform crests (Fig. 17). The saucers are slightly to extremely brecciated although some have horizontal bedding between the thrust pseudoanticlines. Such saucers exist between PA1 and PA2, between PA2 and PA3 and between PA7 and PA8. The spacing between the moderately sized antiforms range from 0.4 m to 1.5 m, whereas the distance between the larger antiforms is between 6 and 15 m. The bases to these structures are not exposed. However, in section H, planar beds are developed beneath the smallest pseudoanticlines. None of the psuedoanticlines are exposed in plan view.

B. LITHOLOGY OF PSEUDOANTICLINES, SALT RIVER AREA

Units A through to C occur in the sections that contain pseudoanticlines and are generally lithologically identical to units A through C in those sections not containing pseudoanticlines. Locally, however, the units are brecciated. Although, the breccias are usually formed of a single unit, mixing of all three units does occur.

UNIT A

Unit A, lithologically identical to unit A exposed in sections H and J, forms a highly deformed traceable marker in the sections that contain pseudoanticlines. The unit has a highly irregular and abrupt lower and upper contact. In places, the unit pinches upward and produces finger-like extensions into the overlying unit B (Appendix 3). In the core of PA4 this unit forms two indistinct antiform

FIGURE 16

16A) Field photograph of two small pseudoanticlines from unit A (section H). The two pseudoanticlines have brecciated cores (a) and are separated by a non-deformed, planar bedded saucer (b). The underlying beds (not shown) are also planar. Note how the beds are upturned on the edges of the crests of the pseudoanticlines.

16B) Field photograph of left pseudoanticline in (A) showing the brecciated core and the upturned adjacent beds (arrows depict dip of beds). A smaller pseudoanticline occurs over the top of this pseudoanticline (a).

16C) Field photograph of pseudoanticlines from unit A (section H). Upturned beds thin across the crest of pseudoanticline (a) and a crestal fracture disrupts the structure (b). Scale bar equals 2 cm.

16D) Field photograph of two small pseudoanticlines from unit A (section H) showing that the brecciated cores (a) are separated by the non-brecciated, planar beds of a saucer (b). Planar beds are developed beneath the pseudoanticlines (c).

16E) Field photograph of a brecciated, chaotically bedded unit A (section H). Calcite cement (white areas) is not forming along joint planes and does extend back into the rock thereby suggesting that the calcite is displacive and brecciating the rock.

16F) Field photograph of unit A. A white, sparry calcite has forced beds of unit A apart (a) with the cement completely surrounding a clast of unit A, thereby having produced a floating texture (b). Large vugs or caves occur where the calcite has been leached (c). Scale bar equals 3 cm.

16G) Field photograph of a large tepee (section E) showing a brecciated core and typical tepee-like crest. Adjacent beds pinch-out laterally over the crest of the tepee (a), beneath planar beds developed above the tepee (b, arrows showing dip of bedding). Fissures containing breccia and shale cut through the planar strata and extend down into the tepee. The bar scale equals 1 m.

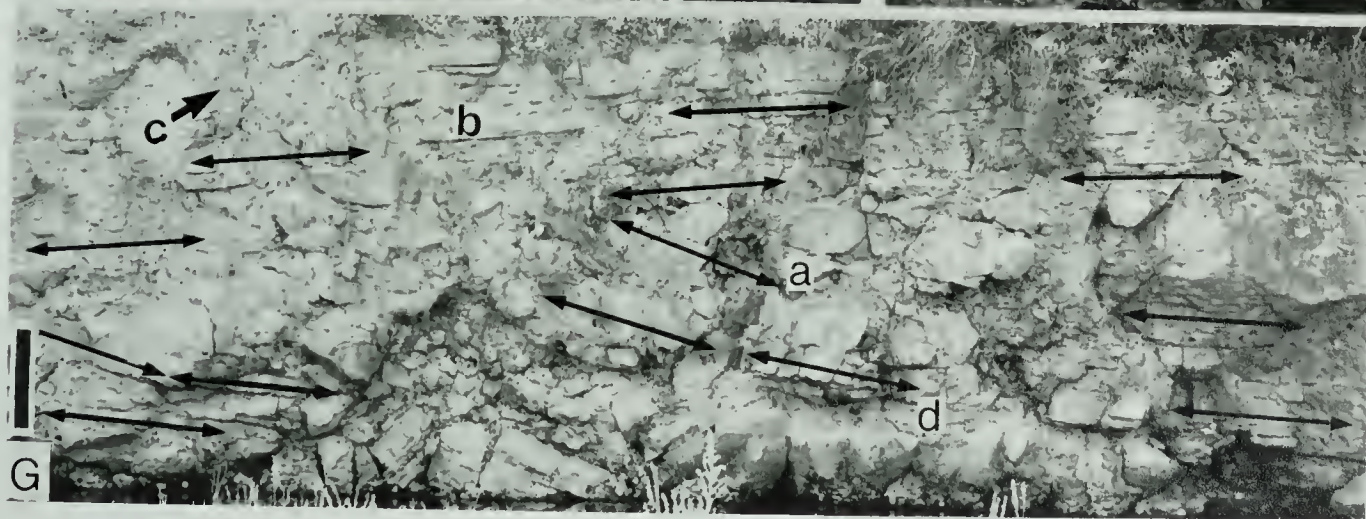
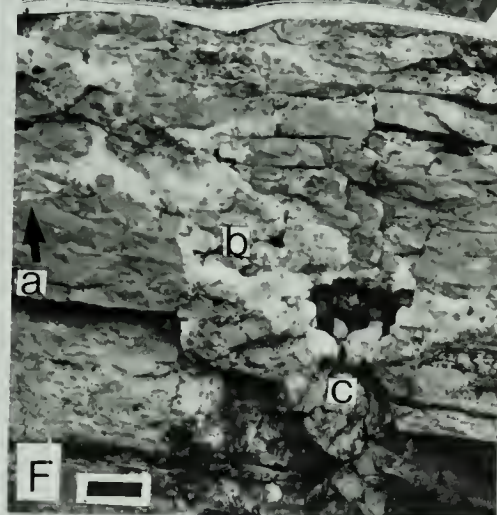
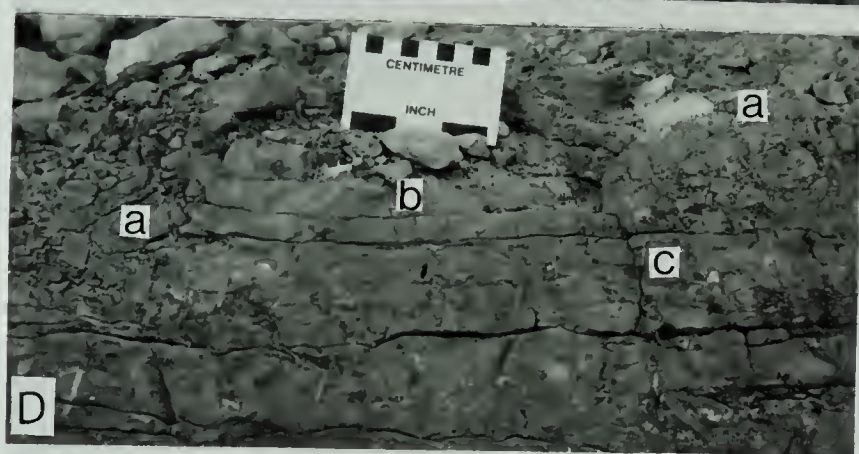
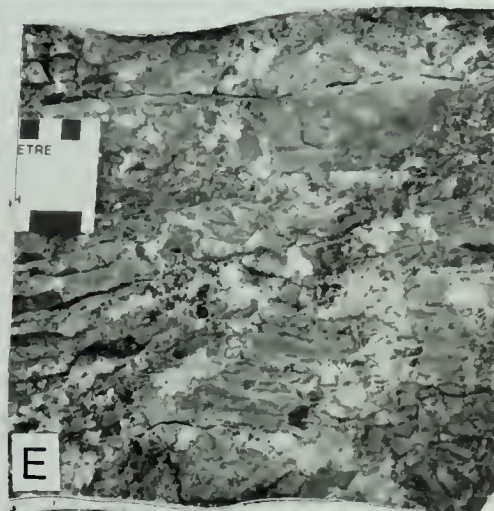
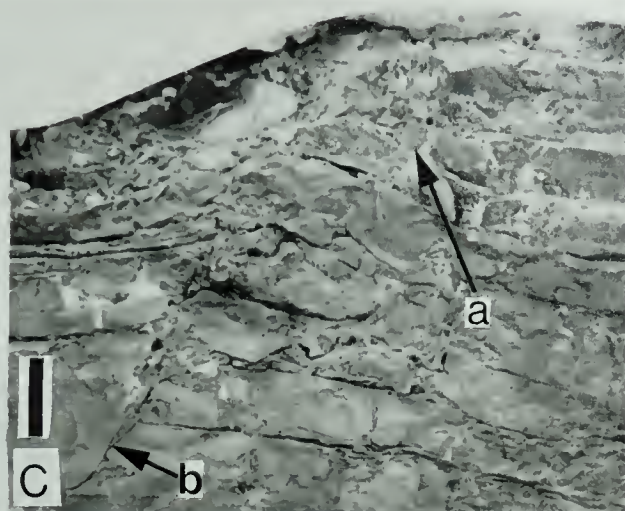
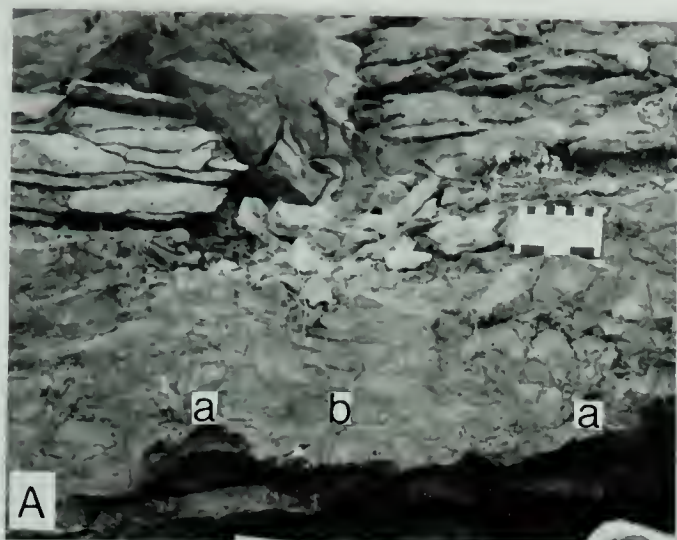
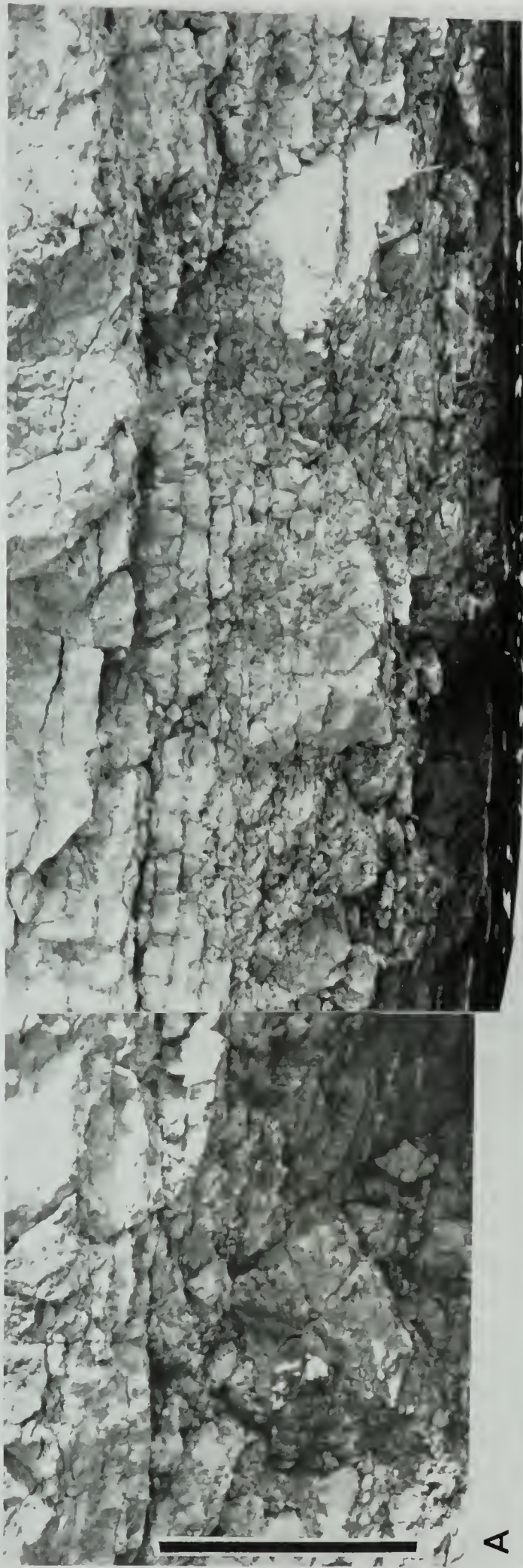


FIGURE 17

17A) Field photograph of section G showing two moderate sized tepees and associated saucer fill. Unit AB is a breccia formed of units A and B. Unit AB underwent extensive dolomitization (stippled area). The unit AB/C1a contact represents a diastem with limestone of unit C1a deposited after the dolomitizing and dedolomitizing events effecting unit AB. PA2 and PA3 have a typical pseudoanticline appearance with beds in the saucer developed between the pseudoanticlines pinched out over the crest of the PA2 pseudoanticlines. Unit C1b is lithologically similar to C1a, however, the abrupt planar contact suggests C1b was deposited as a seperate, major pulse of sediment. Scale bar equals 1 m



A

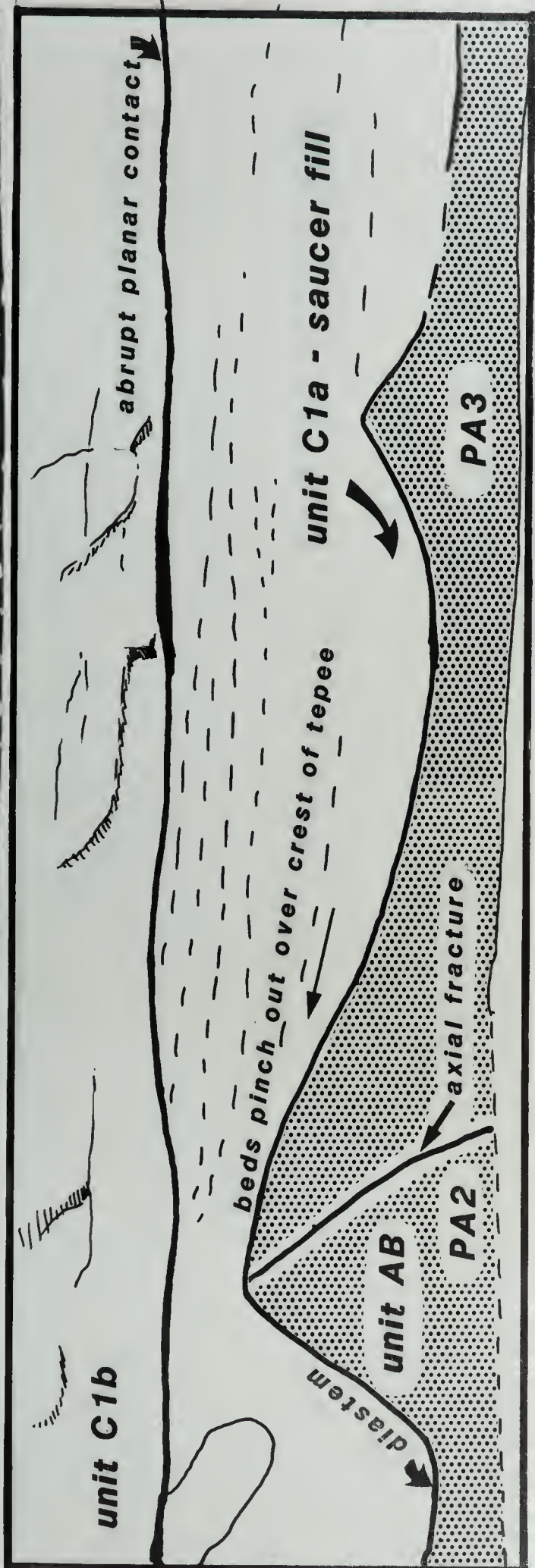


FIGURE 18

18A) Field photograph of large pseudoanticline (tepee) from section G. Unit A forms a black marker in the core of the tepee, with units A and B1 forming a breccia above and below the marker. Units B2/B3 occur as bedded strata over the highly deformed core with an abrupt and slightly wavy contact (a) thus suggesting there is at least two phases of deformation. Scale bar equals 1 m.

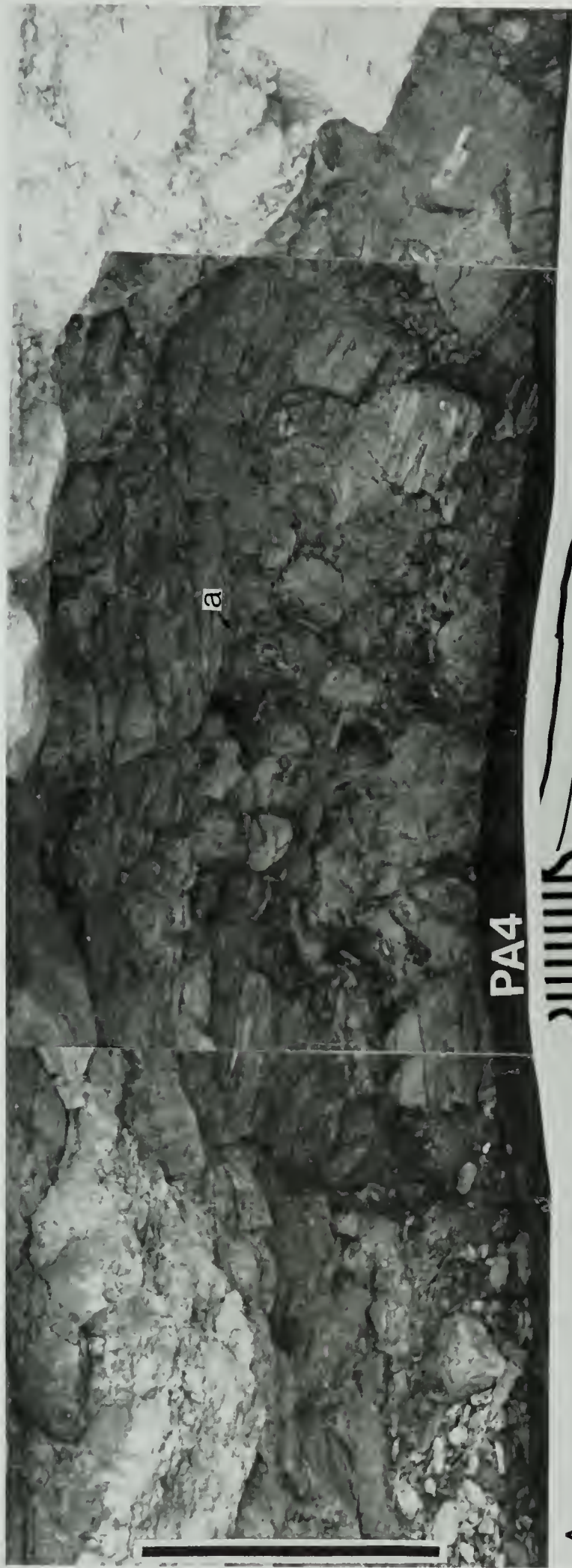


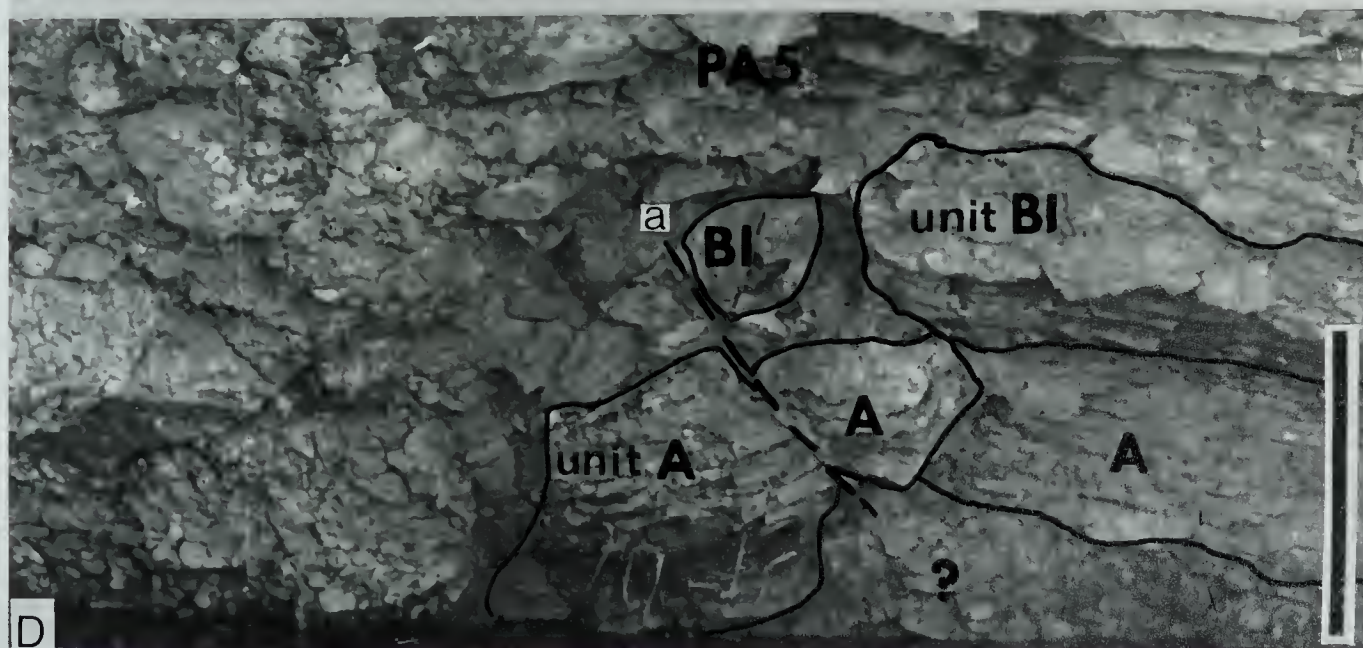
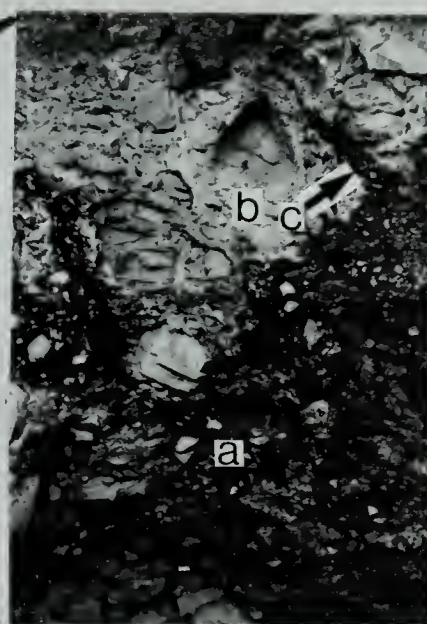
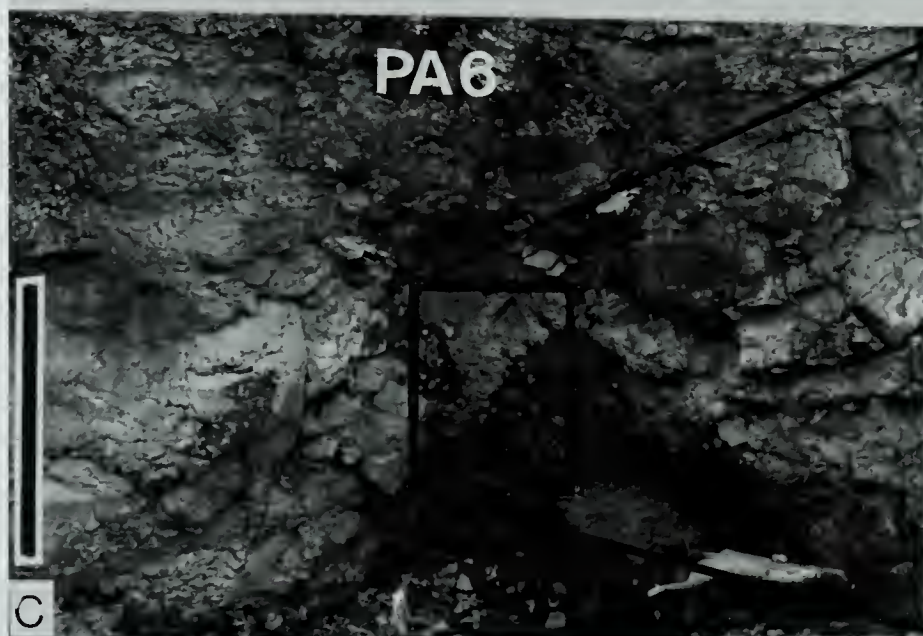
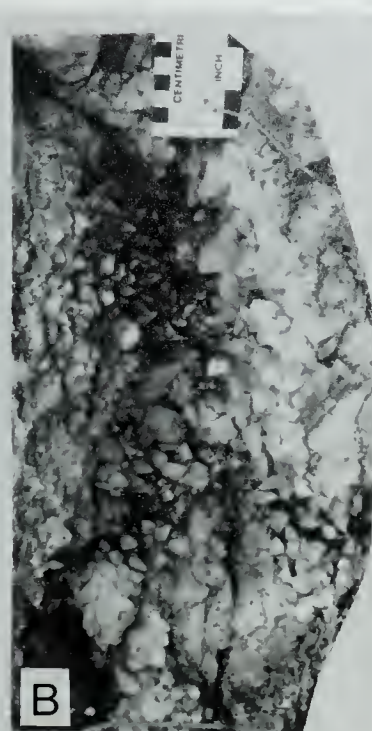
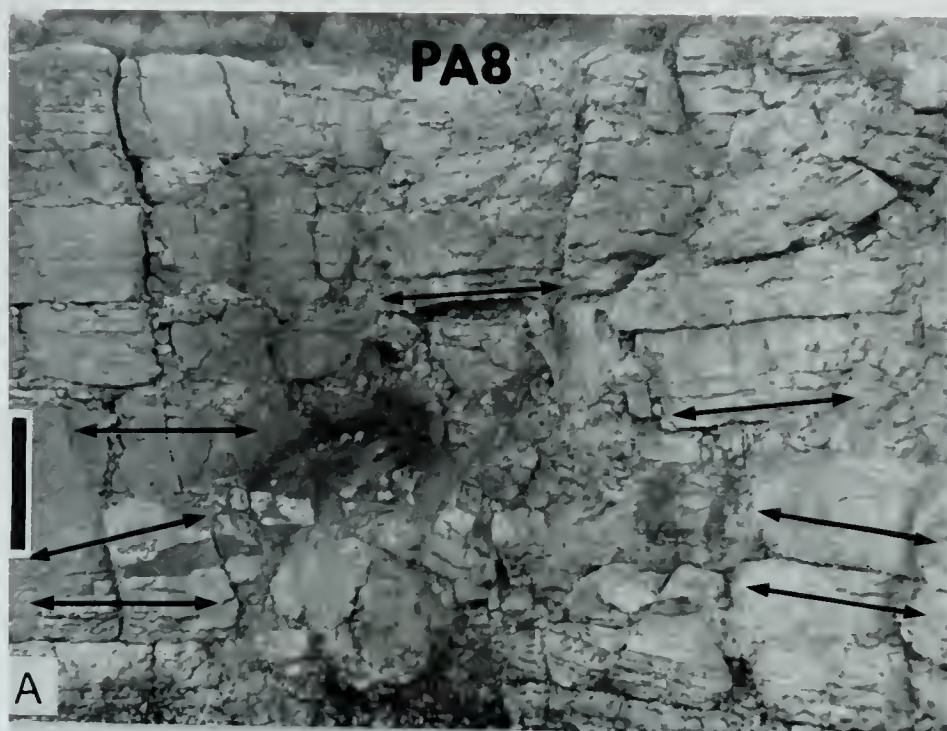
FIGURE 19

19A) Field photograph of a tepee (section E) showing the typical brecciated core and upturned flankbeds. Beds overlying the crest of the tepee are planar and dipping at a different angle (arrows denote dip of bedding) than the underlying beds. Scale bar equals 1 m.

19B) Field photograph of a scalenohedral calcite lined vug (section H). These vugs are common throughout the Salt River strata

19C) Field photograph of a tepee from the south end of section G. The core of the tepee is filled with the lower two fills that comprise carbonaceous, argillaceous breccia. The breccia contains limestone clasts in varying proportions (a). A breccia pipe that is comprised of laminated dolomicrite clasts forms the uppermost fill. The lower fill (a) extends into the breccia pipe (b) as a small fissure (c). This suggests fill (a) was the younger event. Bar scale equals 1.5 m.

19D) Field photograph of tepee from section G showing how a faulted tepee crest (a) disrupts units A and B1. The white dolostone unit beneath unit A is of questionable origin. Scale bar equals 0.5 m.



structures (Fig. 18, Appendix 3). A dolostone unit similar to unit B1 is exposed beneath unit A in section G (Fig. 19D). Tsui (1982) did not describe any dolostone below unit B but did describe a light brown to grey, very fine grained, faintly laminated, 2 m thick argillaceous dolostone to dolomitic limestone that occurred at the base of unit B elsewhere in Wood Buffalo National Park. The strata named unit A could possibly be the basal portion of unit B but Tsui (1982) has correlated this unit with unit A from other sections throughout Wood Buffalo National Park. The underlying dolostone is either additional section of unit B (implying that unit A is not unit A) or the lower portion of unit A has been dolomitized in this area. The latter suggestion is the more likely because dolomitization of unit A in the Salt River exposures reflects supratidal conditions that existed in the Middle Devonian. Such conditions were not reported by Tsui (1982) in other localities in Wood Buffalo National Park where unit A is exposed.

UNIT B

In pseudoanticlines of sections F and G, unit B abruptly overlies unit A. Unit B can be divided into three units that are similar to the three units previously described in unit B. However, in places, unit B3 is difficult to delineate from unit B2. These units display well defined, planar to slightly wavy, bedding. However, local pockets of brecciation commonly disrupt the well defined bedding. In PA4, B2/B3 (B2/B3 implies units B2 and

B3 are not distinguishable from each other) abruptly overlies the highly brecciated core of PA4 (Fig. 18). There, these units are not brecciated, suggesting that they were deposited after the event which deformed units A and B1. Unit B1 is identical to unit B1 of sections H and J. This unit occurs adjacent to the large pseudoanticlines and is represented in large blocks in the core of PA4 (Fig. 18).

UNIT AB

Unit AB is the brecciated equivalent of units A and B. Clasts of unit A and all three divisions of unit B occur in unit AB. Clasts of unit B3 comprise a highly fractured, pale green, finely to medium crystalline, calcareous, dolostone that contains light to dark brown, irregular, vertical to subhorizontal, elongate patches of argillaceous dolostone. These patches represent burrows that are not evident in the other sections. Porosities in B3 can reach 25%, thereby forming an extremely fenestral unit.

Unit AB forms the moderately sized pseudoanticlines at the base of sections F and G. The upper contact with unit C1 is scalloped (Fig. 17) whereas the lower contact is not exposed. As in the other sections, unit AB is extensively dolomitized, whereas the overlying unit C is not. The scalloped contact between units A/B/AB and C marks a period of non-deposition.

UNIT C1a

Unit C1 can be subdivided into units C1a and C1b. In sections F and G these units are easily distinguished, but,

the distinction is not as clear in other sections. C1a overlies unit AB and unit B3 with a sharp, slightly wavy to scalloped contact. Unit C1a appears to wedge out over the crests of PA2 and PA4 (Fig. 17). It is certainly possible that unit C1a was a calcareous sediment that filled saucers developed between the previously formed pseudoanticlines. Unit C1a comprises a semi-prominent, greyish dark brown, dolomitic, very finely crystalline, fossiliferous, rubbly limestone that is, lithologically, considerably different than the underlying dolostone. The unit is highly bioturbated, with the bioturbated areas represented by light brown amorphous patches. The patches, which form 50% of the rock, contain up to 15% dolomite rhombs. Unit C1a contains minor amounts of white rounded anhedral clusters of dolomite that are generally 10 to 20 microns in diameter. These clusters possibly represent centripedal replacement (dedolomite) as described by Shearman et al. (1961). Alternatively, the clusters may simply be evidence of incomplete dolomitization. The latter explanation is more likely due to the fact that the criteria needed to define dedolomites are not present (Shearman et al., 1961). Unit C1a contains bioclastic material that forms up to 30% of the rock, less than 5% ferroan calcite and trace amounts of stylolites, and ferroan dolomite. Eight cm diameter vugs are lined with iron stained calcite in the top 5 m of unit C1a.

UNIT C1b

Unit C1b is separated from unit C1a by a sharp and planar contact (Fig. 17). This contact is well defined since the prominence of unit C1b contrasts sharply with that of unit C1a. The upper contact of unit C1b with the biomicrite marker is gradational over 10 cm. The biomicrite marker extends down into unit C1b via vertically shaped fissures (Appendix 3). However, the irregular contact may be due to weathering characteristics and brecciation related to an event that occurred after the formation of the Salt River antiforms. Unit C1b is similar to unit C1 in the other sections.

The basal 0.5 m of C1b contains abundant solitary corals, colonial corals and crinoid ossicles. It is extensively bioturbated, with up to 15% 50 micron wide amorphous blebs of dolomite and 20 to 30% bioclastic fossil debris. The fossil material is the same as for unit C1. The matrix consists of fossil fragments and smaller clasts of material that are cemented by a clean sparry calcite.

Towards the top of unit C1b there are two distinct calcite sizes. There is an original micrite that has in places been only slightly aggraded and an aggraded very finely crystalline calcite.

UNIT C2

Both the biomicrite marker and unit C2 occur in those sections containing the pseudoanticlines. The lower contact of unit C2 is gradational with the biomicrite marker. The

upper contact is not exposed, but is undoubtedly with the Keg River marker (based on sections elsewhere in the park). Lithologically, the unit is identical to unit C in those sections that do not contain pseudoanticlines. However, in sections that contain pseudoanticlines, unit C2 is extremely fractured and in places extensively brecciated (Appendix 3). Brecciation caused by a later phase of collapse, usually forms in association with the pseudoanticlines.

Subvertical to vertical fissures, filled with brecciated fragments of unit C2, are concentrated above and taper to a point downwards into the pseudoanticlines (Figs. 16G, 19A). Clasts in the fissures generally fine towards the top. A rusty brown to black shale/siltstone is concentrated along the fissures and in pockets in the pseudoanticlines. Rare, 3 cm long acicular crystals of gypsum occur in the shale. Sulphur staining and oxidized iron concretions are common. There is an obvious correlation between the number of fissures and the amount of shale. The greater the number of fissures, the higher the amount of shale.

Vugs are common in unit C2. These vugs, which are up to 0.5 m in diameter, are concentrated in the sections that contain pseudoanticlines (Fig. 19B). They can account for 10 to 20% of the unit and are generally filled with a "dogtooth calcite".

C. CRESTAL DEFORMATION AND SEDIMENTARY FILL OF PSEUDOANTICLINES

Crestal deformation of the pseudoanticlines, generally in the form of thrust faulted structures, occur in PA2, PA3 and PA5 (Figs. 17, 19D, Appendix 3). In unit AB, clasts of unit B3 are extensively fractured and brecciated with argillaceous lined stylolites outlining the edges of the large fractures. Clast compositions filling the fractures include, bioclastic limestone, slightly dolomitized bioclastic limestone, organic opaques, fossil fragments, shale and minor amounts of gypsum and dolostone. The dolostone is derived from the fractured host unit B3, implying that the fracturing occurred after dolomitization. Evidence of post-dolostone fracturing is given where micron sized dolomite seams from unit B3 are contained in clasts in the fractures. The bioclastic limestone is derived from unit C and is not extensively dolomitized. These limestone clasts were originally only partially lithified as they fell into the fractures for some are deformed around the relatively older, more lithified, dolostone clasts. Partial lithification of the bioclastic limestone was probably due to early aragonite or high magnesium calcite cementation. Individual, highly abraded fossil fragments in the fractures suggest that sediment, not rock, filled the fractures.

In PA4, the matrix consists of clasts of all three units (units A, B and C), as well as pelmicrites that contain argillaceous laminae and pellets rimmed with fibrous

calcite cements. Fibrous cement and organic opaques also rim clasts comprised of horizontally laminated, finely crystalline to medium crystalline, dolostone and shale. Pure, featureless, porous dolostone and dark carbonaceous shale occur at the base and in pockets throughout the core, respectively.

In PA6, the core of the pseudoanticline is filled with three different rock types (Fig. 19C) The lowermost fill is formed of a mixture of poorly indurated shaly siltstone and limestone clasts. The argillaceous matrix is light brown, calcareous, carbonaceous and extremely silty. The limestone clasts, which form less than 5% of the unit, are angular, light grey to dark brown, medium crystalline limestone that consist of laminated layers of 30 micron wide anhedral calcite. Highly argillaceous and non argillaceous calcite layers form interlamination in the limestone clasts. The calcite in the argillaceous layers is a finer crystal size than calcite in the non argillaceous layers. Laminae tend to "fade" internally, reflecting the gradational lowering of argillaceous material. The gradational argillaceous content of laminae, the evident rippling and a hint of cross-lamination in the clasts, suggest that deposition occurred under waning water conditions.

Overlying the lower fill is a zone that contains a higher percentage of clasts. The clasts, which are up to 7 cm along the longest axis, form 30% of the fill. Clasts under 1 cm comprise 70% of all the clasts. This fill is more

argillaceous, but less carbonaceous than the lower fill. The clasts comprise breccia that are, in turn, comprised of fragments of shale, limestone, dolomite and dedolomite. Crystal size ranges from aphanocrystalline to medium crystalline. Less than 5% of the aggraded calcite is replaced by silica. There is a high organic content in both fills. Porosity is up to 30 to 35% of the rock.

The third type of fill is a white dolostone breccia that forms a funnel-like pipe at the top of the core in PA6 (Fig. 19C). This unit is comprised entirely of a brecciated, very porous (up to 35%), buff, very finely crystalline dolostone. Subangular to angular, dolostone blocks have light to medium brown laminations that are randomly oriented from block to block. The matrix comprises a fine white chalky powder. The matrix to clast ratio is approximately 50:50. The dolostone fill becomes intermixed with the other fills towards the base of the core (Fig. 19C). The second fill extends into the dolostone pipe near the top of the core suggesting that it was a later event than the pipe. However, since preserved fossils were not found, the material in the core could not be dated. Grey oxidized crusts have formed on the surface of the fills, thereby suggesting that groundwater has at one time percolated through the units.

This sedimentary fill can be interpreted in two ways. It could be either a penecontemporaneous sedimentary fill which was falling into a crestal fracture while the

pseudoanticline was forming, or it was the result of subsurface solution and subsequent infill of the pseudoanticline sometime after its formation and burial.

D. SUBAERIAL EXPOSURE AND PENECONTEMPORANEOUS FORMATION OF THE KEG RIVER FORMATION PSEUDOANTICLINES

Pseudoanticline deformation along the Salt River began with deposition of unit A. The small pseudoanticlines in sections G and H (unit A) probably represent deformation of pre-lithified crusts in a subtidal setting. The abundance of calcite cement suggests that the force of crystallization of the cement was the main deforming agent. Planar beds developed at the base of the smaller pseudoanticlines suggest that they were not all the result of a single episode of deformation. The beds deposited over the previously formed pseudoanticline are horizontal suggesting that the two events were separate. The later sediment was also deformed, but the later pseudoanticlines do not vertically coincide with pseudoanticlines produced during the first phase of deformation (Figs. 16B, 16E). Unit A therefore represents repeated cycles of deposition and deformation.

During regression of the Devonian seas dolomitization occurred, followed by dedolomitization and lateral expansion of a partially lithified crust in a supratidal environment and the formation of the two larger pseudoanticline types.

The diagenetic changes and deformation mark a period of subaerial exposure and emergence of unit B throughout the study area. Evidence suggests that the top of unit B3 and possibly the top of unit B1 mark periods of non-deposition. In the core of PA4, unit A and unit B1 are extensively brecciated with unit A outlining indistinct, brecciated antiformal structures (Fig. 18, Appendix 3). Unit B2/B3 lies above the deformed core with an abrupt and slightly wavy contact. Clearly, units A and B1 were deformed prior to deposition of units B2/B3, thus suggesting that the contact between unit B1 and units B2/B3 marks a period of non-deposition (Fig. 18). However, this interval of non-deposition is not apparent in the northern part of section F where units A and B are deformed in highly brecciated pseudoanticlines (unit AB).

In the moderate and large sized pseudoanticlines non-deposition and subaerial exposure also occurred after deposition of unit B3 because the supratidal rocks of unit B were brecciated into unit AB prior to deposition of unit C1. It is evident that unit AB was deformed after the dolomitization, because fractures in the unit are filled with a bioclastic limestone of unit C1a. This suggests that there was a period of non-deposition between the dolomitization and the deformation, during which time the undolomitized material fell into the fractures. The deformation represents formation of a pseudoanticline, with the period of non-deposition marking a period of subaerial

exposure. The time of exposure between deposition of unit B3 and unit C1 was the minimum time needed to dolomitize and dedolomitize the sediment of units A and B in a supratidal environment. This observation is in agreement with subaerial exposure postulated at the end of unit B3 deposition in the other sections.

After formation of the pseudoanticlines, unit C1a formed the basal deposit of a transgressive sequence that filled the saucers between the pseudoanticlines (Fig. 17, Appendix 3). The intertidal, highly bioturbated beds of C1 thin over the top of PA2 (Fig. 17). Unit C1b forms the distinctive abrupt capping unit over the top of the pseudoanticlines (Fig. 17, Appendix 3).

In summary, the history of the sections containing the pseudoanticlines is as follows:

1. Unit A was deposited, followed by development of the small scale psuedoanticlines (section H).
2. Unit B1 was deposited. Regression occurred, leading to the subaerial exposure of unit B.
3. Larger pseudoanticlines began to form (core PA4). Units A and B1 were deformed into pseudoanticlines in the core of PA4. The larger pseudoanticlines represent large megapolygonal domed areas that were forming along an ancient coastal shoreline.
4. A period of non-deposition followed deformation of units A and B1. A planar, slightly wavy, horizontally bedded unit B2/B3 was deposited over the top of the deformed

core PA4. This suggests that different stages of pseudoanticline deformation occurred in the Salt River strata, with some pseudoanticlines being composite structures.

5. Just prior to subaerial exposure extensive dolomitization occurred in a "mixing zone" environment. Rare gypsum crystals replaced phase 1 and phase 2 dolomite and were later dissolved and filled with phase 3 dolomite.
6. During continued regression marine waters changed to fresh water as the environment representative of the Salt River strata migrated from a marine setting through to a subaerially exposed environment. The regression was marked by an increasing input of fresh water that eventually led to dedolomitization and synchronous aggradation of remaining calcite. Diagenesis was extensive and considerably altered the original sediment.
7. The large pseudoanticlines continued to deform.
8. A diastem occurs between units B3 and C1a.

The moderately sized pseudoanticlines formed (PA1, PA2, PA3, PA5) during this period of nondeposition. These pseudoanticlines became brecciated after deposition of unit B3 because unit A and B form their brecciated equivalent, unit AB. These pseudoanticlines are smaller structures that were superimposed on the larger pseudoanticlines. The extensively dolomitized and

dedolomitized units A and B, were fractured during deformation. The fractures were later filled with a nondolomitized to slightly dolomitized bioclastic limestone (unit C1a). A diastem exists between the dolomitizing/dedolomitizing events and the fracture filling event.

9. The intertidal to subtidal sediments of unit C1a filled the saucers that existed between the crests of the pseudoanticlines. Unit C1a pinches out against the larger pseudoanticlines, which were still deforming.
10. Deposition of unit C1b with a sharp and planar lower contact. There was minor pseudoanticlinal deformation of this unit in section G (PA4). By comparison, pseudoanticlinal deformation of this unit in section E is obvious. In section E, the pseudoanticlines continued to grow up into unit C2.
11. Deposition of the biomicrite marker and unit C2. The assumption that pseudoanticlinal growth stopped midway through unit C2 is supported by the fact that the uppermost strata of unit C2 form planar, undeformed beds over the top of the pseudoanticlines (Figs. 16E, 19A).

E. GENETIC THEORIES FOR THE SALT RIVER ANTIFORMS

PREVIOUS THEORIES FOR THE SALT RIVER ANTIFORMS

Tsui (1982) outlined four possible mechanisms for the formation of the antiform structures that occur along Salt River, namely;

1. ice thrusting
2. gypsum and anhydrite intrusions
3. salt intrusion
4. hydration of anhydrite to gypsum

Ice thrusting produces sharp crestral ridges that have fold axis trending parallel or perpendicular to the glacial front (Tsui, 1982). They have been described as structures with a height of 61 m and an average basal width of 183 m. Based on glacial striations in the Salt River Area the glacier ice flow had a 222° trend (Tsui, 1982). Tsui (1982) suggested that this was an unlikely mode of formation for the Salt River antiforms because the trends of the antiforms are oblique to the ice flow trend and the antiforms are nowhere near the dimensions of previously reported ice thrust antiforms.

Tsui (1982) favored the hydration of anhydrite to gypsum theory, stating that "Antiform structures discovered near the South Salt River Bridge area are formed by the hydration and expansion of anhydrite in the Chinchaga Formation. This produced gypsum diapirs and resulted in uplifting and deformation of the beds above" (Tsui, 1982, p. 107). These structures were related to normal faults that had formed post-Devonian and possibly just pre-glacial (Tsui, 1982). Many of the larger antiform structures studied by Tsui (1982) to the north of this study area possibly formed in relation to post-Devonian faulting, however, the antiforms studied in this paper did not. These structures

actually formed by different processes during the Middle Devonian.

Although the Cold Lake and Chinchaga Formations underlie the Keg River Formation in this area, salt intrusions, gypsum and anhydrite intrusions and the hydration of anhydrite to gypsum (producing gypsum diapirs) are not feasible genetic theories for the Salt River antiforms. Intrusions of evaporites generally refer to large domal structures that have diameters on a kilometer scale. Tsui (1982) stated that evaporite domes can create anticlinal structures with a diameter of at least 1.6 km. The Salt River antiforms, however, are only a few meters in width. The 50 cm (max.) wide pseudoanticlines that occur in section H (unit A) have planar beds developed below them (Fig. 16C, 16D). If these structures were formed due to any type of evaporitic intrusion, or diapir, the underlying beds would also have to be deformed. This is certainly not the case for those beds underlying the 50 cm wide pseudoanticlines in section H. Also, a doming or intrusive event could not form the two to three generations of growth evident in the larger composite pseudoanticlines of sections G and H.

There are various reasons that suggest that these pseudoanticlines formed during the Devonian. The existence of a diastem at the top of unit B, subsequent extensive diagenetic changes of rocks below the diastem and not above the diastem, subaerial exposure beneath the diastem, and

deformation prior to deposition of unit C1 all support this contention. The evidence suggests, however, that some antiforms in the Keg River Formation formed during, as well as prior to, deposition of unit C. Nevertheless, pseudoanticline growth was terminated in the middle of unit C2 because bedded strata in the upper portion of unit C2 truncate bedding of the pseudoanticline forming strata. This suggests that the strata was deposited after the pseudoanticline formed (Figs. 16G and 19A).

TERMINOLOGY

The Salt River antiforms are considered to be pseudoanticlines, however, there are different types of pseudoanticlines. Therefore, in order to interpret the Salt River antiforms, there must be an understanding of the different types.

Pseudoanticlines were first reported in a caliche horizon at shallow depths below the surface of Holocene deposits of Central Tamaulipas, Mexico (Price, 1925). These pseudoanticlines involved buckling of strata due to caliche growth. Similar structures in caliche horizons have been termed expansion structures (Bretz and Horberg, 1949; Giles et al., 1966) or buckle cracks (Reeves, 1970).

"Pseudoanticline" is not to be confused with the term anticline, which is generally considered to be a convex fold with the oldest rocks occurring in the center of the fold. The latter is a result of tectonic plate interaction whereas the former is not. Jennings and Sweeting (1961, p. 635)

suggested that "It may be best to retain a broad connotation for 'pseudoanticline' (cf. German Pseudo-dislokation) and qualify it for specific types...".

The term tepee represents one specific type of pseudoanticline, namely that structure which resembles a cross-section of an American Indian tent (Adams and Frenzel, 1950). These structures occur in a back barrier facies of the Permian Carlsbad Group in the Guadalupe Mountains. However, they have also been documented in peritidal environments (Davis, 1970; Assereto and Kendall, 1971, 1977; Evamy, 1973).

Based on the spatial relationship of the pseudoanticline with respect to the depositional/diagenetic environment (i.e. subtidal, intertidal-supratidal, supratidal-continental) Assereto and Kendall (1977) defined the following specific types of non-tectonic pseudoanticlinal structures.

1. Submarine Pseudoanticlines - anticlines forming at the edges of expanded subtidal megapolygons formed in carbonate beds during early submarine cementation (Assereto and Kendall, 1977).
2. Tepee - antiform structures that form at the margins of near marine megapolygons; developed by penecontemporaneous expansion of peritidal crusts: embryo, mature and senile stages can be recognized (Assereto and Kendall; 1977).
3. Caliche pseudoanticlines - the saucer-like structures

that form by expansion of calcareous crusts within a fresh-water soil profile (Price, 1925).

GENESIS OF PSEUDOANTICLINES

Assereto and Kendall (1977) listed six hypotheses for the formation of pseudoanticlines, stating that there was a general agreement among authors as to the origin of submarine and caliche pseudo-anticlines (Shinn, 1969; Dunham, 1962; Jennings and Sweeting, 1961; Reeves, 1970). The origins of tepee structures, however, are more varied.

GENESIS OF SUBMARINE PSEUDOANTICLINES

There is only one process that can be applied to the genesis of a pseudoanticline in a subtidal environment, namely crystallization forces associated with the growth of aragonite cement. Precipitation of aragonite crystals causes a lateral expansion of a partially cemented/lithified crust. The lateral compressional forces are relieved when deformation occurs, resulting in a submarine pseudoanticline. Lateral movement is confirmed by the presence of slicken-sides on the lower surfaces of some cemented slabs (Shinn, 1969).

GENESIS OF CALICHE PSEUDOANTICLINES

Caliche pseudoanticlines are also the result of expansion processes (Jennings and Sweeting, 1961; Reeves, 1970; Price, 1925; Watts, 1976; Assereto and Kendall, 1977). General consensus for the type of "expansion" responsible for a caliche pseudoanticline

suggests that the force of crystallization of calcium carbonate in a plugged, mature, caliche horizon is the most important factor. Aristarain (1969) and Allen (1974) stated that the displacive force of crystallization is recorded by the splitting apart of quartz grains in Quaternary and ancient caliche. Some of the pseudoanticlines in South Africa did not result from horizontal expansion of the sediment, but by partial collapse of the calcrete due to the removal of underlying salts (Watts, 1976). Swelling clays (Watts, 1976), moisture swelling and thermal contraction-expansion (Assereto and Kendall, 1977) may add to the deformation process.

GENESIS OF TEPEES

There are many theories pertaining to the formation of tepees. Adams and Frenzel (1950) and Boyd (1958) suggested tepees formed as a result of tectonic forces. However, Assereto and Kendall (1977) discounted this theory by showing that the expansion and brecciation of the tepees, along with their geometric symmetry (in plan view) argued against such an origin.

The expansion hypothesis is the most widely accepted hypothesis. This hypothesis includes a wide variety of modes of "expansion" associated with pseudoanticlines. The formation of a tepee is the result of a combination of the following modes of expansion: evaporite hydration and dehydration, thermal expansion,

moisture swelling, force of crystallization, contraction and expansion due to plant factors (Assereto and Kendall, 1977).

Hydration of anhydrite to gypsum is accompanied by a 30 to 40% volume increase (Deer et al., 1977 ; Pettijohn, 1949), thereby applying high stresses on surrounding rocks (Assereto and Kendall, 1977). For example the transition of anhydrite to gypsum caused violent explosions and uplifts in Texas (Brune, 1965) and uplift and deformation in Manitoba and northeastern Alberta (Bannatyne, 1959). Such deformation can occur quickly; for example, concrete linings of a trench in northern Italy were deformed and fractured in a period of ten years (Redfield, 1963). The development of polygons of anhydrite just below the surface of supratidal sediment in the southwest Persian Gulf (Kendall and Skipworth, 1969) and dehydrated salts that are wetted and rapidly hydrated, thereby causing stress on fissure walls (Assereto and Kendall, 1977), are examples of disruptive growth of evaporites in Holocene examples. Tepees in the Permian Reef Complex may have resulted from expansion associated with the anhydrite to gypsum transition (Newall et al., 1953). The importance of crystal expansion of evaporites, with regard to the development of tepees, is not well known (Assereto and Kendall, 1977). However, since evaporites are highly soluble they are not common in ancient tepees, therefore

are not considered a major expansion mode.

Displacive carbonate, due to thermal expansion and contraction of sediment in supratidal flats, can form tepees (Evamy, 1973; Burri et al., 1973). Assereto and Kendall (1977) suggested that the average linear thermal coefficient for carbonate rocks does not substantiate the large expansion of carbonate needed to form tepee structures. In a 10 m wide polygonal saucer a temperature increase of 50°C can produce a linear expansion of about 4 mm. "Minor" superficial fractures in embryo megapolygons may originate by thermal expansion (Assereto and Kendall, 1977). The magnitude of moisture swelling cannot cause extensive fracturing in carbonates (Goudie, 1974); however, the process may be responsible for enlarging previously formed fractures in the carbonate.

Growth of calcium carbonate crystals is thought to be a displacive force in carbonate rocks. A supersaturated calcite medium can produce crystals of calcite that exert stresses of ten atmospheres (Weyl, 1959). Assereto and Kendall (1971, 1977) and Smith (1974) suggested that tepees formed due to expansion of crystallizing cements in internal pores, joints and fractures.

One of the major arguments in favour of "force of crystallization", in ancient examples, is the presence of "floating textures". However, the validity of

floating textures is questionable. Floating textures are textures where a piece of the rock has been "ripped away" from the mother rock and completely surrounded by cement. In two dimensions the displaced clast appears to be floating, however, the clast may have been supported in the third dimension. Watts (1978) suggested that the growth of displacive micrite can forceably brecciate rocks. "Although I would not insist that displacive calcite is common in all calcretes . . . and I find no reason for doubting the existence, or importance of displacive calcite growth within some calcretes" (Watts, 1979, p. 422). Ferguson et al. (1982) and Warren (1982) suggested that cements are important in the formation of tepee structures. However, they argued that the cements do not cause the tepee structures due to "displacive crystallization", but by playing a much more passive role in the genesis of the tepee. They are simply precipitated into a void space in the lithified rock with environmental changes causing the actual deformation.

Klappa (1978) suggested that floating textures in caliche from eastern Spain, from Barbados (James, 1972; Harrison, 1977), from Shark Bay, Australia (Read, 1974) and from northern Tanzania (Hay and Reader, 1978) can be explained by mechanisms other than displacive calcite. From these examples Klappa (1979) stated there is no conclusive evidence to indicate displacive calcite

is responsible for floating textures.

In a Quaternary calcrete from eastern Spain, tepee structures are formed due to a mechanical break up of the rock by plant roots (Klappa, 1978). According to Klappa (1978,) the roots allow carbonate rich fluids to migrate into the disrupted rock and recement the breccia. This is a passive process and the resultant tepee is not the product of displacive calcite. In contrast, Ferguson et al. (1982) and Warren (1982) reported plants that selectively grow along tepee crests in South Australia. However, they stated that the plants do not have any genetic relationship with tepee formation. Plant roots in tepees of Tamaulipas, Mexico simply assisted in opening joints, which in turn assisted in the development of caliche and subsequent deformation (Price, 1925). Watts (1979) stated that root activity does cause some brecciation, but it is not significant in pseudoanticline formation. The above evidence suggests plants play a minor role in tepee development.

"Dessication contraction is thought to play an important role in the development of peritidal tepees by determining their polygonal pattern" (Assereto and Kendall, 1977, p. 198). Intense evaporation beneath the surface of a lithified crust may be important in the formation of dessication fractures in tepees. Kendall (1969) and Evamy (1973) further suggested that tepees

were formed partly by dessication. However, if tepees were formed by dessication then the fractures would be wide at the top, taper vertically downward and the steepest beds would be near the edge of the fracture (Smith, 1974). However, in tepees, the beds near the crest are generally flat to arched over and if the criterion above do not occur, Smith (1974) concluded that the tepee does not have a dessication origin.

The multiple genetic hypothesis has tepees developing due to a continual repetition of the following processes (Assereto and Kendall, 1977):

1. Dessication and thermal contraction cracks initiated during low water levels;
2. Periodic wetting causing enlargement of the fractures by moisture swelling;
3. Further enlargement by the displacive force of crystallization by speleothemic cements. These cements cause the major deformation of the tepee.
4. Hydration of minerals, thermal expansion and faulting add to the disruption of the sediment.

A better genetic hypothesis for the formation of pseudoanticlines was proposed by Warren (1982) for tepees that occur in coastal salinas in South Australia. Tepees resulted from hydrologic seasonal changes that effected partially lithified crusts on the edges of salinas. Ferguson et al. (1982) reported a similar groundwater upwelling process in the Fisherman Bay area,

South Australia. The "Salina hydrology not only controls the formation of the veneer boundstone (Warren, 1982 used the term boundstone as an alternate term for pellitoidal mudstone, pelleted grainstone and recrystallized limestone), it also controls the formation of the tepees and mound structures found in the veneer" (Warren, 1982, p. 1186). Seasonal ground water fluctuations cause expansion and subsequent contraction of a pre-cemented crust, which in turn governs the formation and shape of the tepees. When the pore pressures are high the pre-cemented crust is domed, thereby causing tensionally formed triradiate polygonal fractures. Any pre-existing tepees become flatter and the dips of the flanking beds change, thus showing the tepees to be dynamic, not static structures. The tensionally formed cracks are sights of preferential groundwater upwelling that fill with sediment and/or aragonite, thus adding to the total volume of the cracked lithified crust (Ferguson et al., 1982; Warren, 1982). Since pore pressures decline, the lithified crust collapses under its own weight. Since the crust had increased in size, due to the addition of sediments and cement, it undergoes deformation when the annual increment in size is taken up by the overthrusting tepee structures.

Warren (1982) stated that these tepees could not be explained by previous outlined hypotheses. Therefore, he

suggested such tepees could be termed "boxwork or groundwater tepees". The yearly repetition of the following processes generates groundwater tepees (Warren, 1982).

1. A rigid crust forms at or near the groundwater table in association with schizohaline, brackish to hypersaline waters.
2. Pore pressure builds up in the boxwork boundstone until both this unit and the overlying veneer boundstone expand elastically thus producing triradiate polygonal, formed cracks.
3. The lithified crust is bathed in a brine supersaturated with aragonite. Aragonitic speleothemic deposits are precipitated in fenestrae and large cavities within the crust. Neomorphic aggradation also occurs.
4. Pore pressures are depleted causing subsequent collapse. The crust is compressed with subsequent overthrusting taken up by the triradiate megapolygonal cracks. The crust is subaerially exposed with salt weathering and thermal expansion breaking it into fragments. These fragments are cemented to form a new surface of the crust.

The influence of groundwater on lithified crusts on the sea floor is the best genetic hypothesis for the formation of tepees to date.

CLASSIFICATION OF TEPEES - A CLASSICAL APPROACH

Assereto and Kendall (1977) defined three types of pseudoanticlines which represent "growth stages" within a gradational spectrum. Thus, the formation of a pseudoanticline can be traced from its origin, a submarine pseudoanticline, to its end, a caliche pseudoanticline. A pseudoanticline migrates through time from a "wet" submarine stage to a high and dry stage, generally represented by a caliche environment. The variation in shape and complexity of the dynamic, ever changing, pseudoanticline allows an easy distinction between "wet and dry" end members; but, distinction among the three tepee types is more difficult (Assereto and Kendall, 1977).

The transition in the tepee spectrum, from the embryo stage to the senile stage, represents an increasing complexity of morphology, cements and exposure time (Assereto and Kendall, 1977). Time intervals between the formation of fractures and cavities and the subsequent filling of the void spaces with sediments and cements becomes larger and more complex as the tepee structures mature.

The embryo tepees form in a low supratidal environment where they begin as thin fenestral crusts formed by precipitation of calcium carbonate in the vadose zone (Assereto and Kendall, 1977). Mature tepees occupy a higher position on the supratidal flat, which results in thicker crusts, larger polygons and higher evaporation rates.

Concretionary pisolites, which form as a result of aragonite precipitation in the schizohaline waters of the vadose zone, are common in the mature form (Assereto and Kendall, 1977). Mature tepees are exposed above the high water mark longer than embryo tepees and speleothemic deposits become proportionally more complex and abundant with increasing time of subaerial exposure. Therefore, the oldest tepee structures (senile tepees) have higher amounts of cement, with the cement types and textures being more exotic. Periods of exposure are frequently interrupted by marine flooding (eg. storm generated tides) with the frequency of interruption decreasing as the tepee gets older and migrates away from the marine environment. It is this periodic flooding that causes the variety of exotic cements and cross-bedded calcarenites that fill fractures and cavities of the older tepees (Assereto and Kendall, 1977). Fluctuations in the water level, caused by seasonal discharge of water from Pleistocene dune aquifers, is responsible for several additions of cement to the tepee structures in South Australia (Warren et al., 1982). Ferguson et al., (1982) suggested similar reasons for a fluctuating water table in the Fishermans Bay Area, South Australia.

Senile tepees form from mature tepees that are exposed to long periods of supratidal or continental conditions (Assereto and Kendall, 1977). Senile tepees, the most complex tepee type, have a highly porous fabric that

suggests that the tepees have undergone extreme vadose dissolution. So extreme that a rock can consist of less than 20% original material. Warren (1982) described senile tepees as consisting of intraclast breccia fragments in association with karstic dissolution.

GROUNDWATER TEPEE

Assereto and Kendall (1977) assumed that tepees form in a peritidal environment. This assumption appears valid for the "classical" concept of tidal related tepees (eg. subtidal, intertidal, supratidal to continental). However, tepee structures have formed in coastal salinas, devoid of any relationship to tidal action (Warren, 1982). This pseudoanticline form undergoes periodic intervals of exposure and wetting, similar to the classical concept of a tepee, thus warranting the name tepee. However, this structure does not fall into the original definition of a tepee, for is not formed in a tidal environment or by the same means. This form of tepee does not fit into the evolving pseudoanticline scheme, which is the basis for the "classical" tepee classification. For these reasons, the suggestion made by Warren (1982) in forming a new tepee form is followed. The term groundwater tepee (Warren, 1982) is an appropriate term.

F. COMPARISON AND CLASSIFICATION OF THE KEG RIVER PSEUDOANTICLINES

PSEUDOANTICLINES OF UNIT A

The small pseudoanticlines which occur in unit A, sections F, G and H are very similar to submarine pseudoanticlines.

MORPHOLOGY OF SUBMARINE PSEUDOANTICLINES

Of all the pseudoanticline structures, submarine pseudoanticlines are the least reported. The only Holocene examples are those in subtidal waters of the Persian Gulf (Shinn, 1969; Kendall and Skipworth, 1969). Submarine pseudoanticlines occur as polygonal patterns that extend over the sea floor for tens of square miles, in water 2 to 19 m deep (Shinn, 1969). Submarine polygons, 10 to 40 m wide, generally have sharp and regular polygonal patterns (Assereto and Kendall, 1977).

The only ancient example of submarine pseudoanticlines, as designated by Assereto and Kendall (1977), was documented by Lindstrom (1963). The Ordovician submarine pseudoanticlines of Sweden are gentle to overthrust folds that are generally 0.5 m wide and 15 cm high.

Plan views of the Salt River pseudoanticlines are not evident because of glacial overburden. The pseudoanticlines in unit A are nowhere near the dimensions as suggested by Assereto and Kendall (1977); however, they are similar in size to the submarine

pseudoanticlines described by Lindstrom (1963).

The ridges that constitute the polygonal structures are thrust up over one another, thus forming severely fractured antiforms in cross-section (Shinn, 1969). These ridges comprise lithified sediment (beds) that underwent early cementation in a submarine setting. The lithified beds are interbedded with unconsolidated, loose sediment. Generally groups of 2 to 3 beds occur, with individual beds being 5 to 10 cm thick. In unit A, of the Salt River strata, individual beds are 5 cm thick.

Like submarine pseudoanticlines documented elsewhere (Assereto and Kendall, 1977), the pseudoanticlines occurring in unit A, along the Salt River, also display flat and eroded crests (Fig. 16A, 16B).

LITHOLOGY OF SUBMARINE PSEUDOANTICLINES

Strata in submarine pseudoanticlines generally consists of allochemical sediment (Assereto and Kendall, 1977), micrite (Lindstrom, 1963) and pelleted carbonate muds (Shinn, 1969), that are usually partially lithified with an early cement. Unit A is lithologically similar to the subtidal sediment reported by Shinn (1969).

The lack of diagnostic features of intertidal and supratidal environments can support the contention of a submarine lithology. Features such as particle orientation, birds eye vugs, inclined or horizontal

lamination (Shinn, 1969), fenestrae and pisolitic textures (Assereto and Kendall, 1977) are not present in submarine pseudoanticlines. Of the above textures only horizontal lamination is observed in the smaller pseudoanticlines located along the Salt River.

Replacement textures and recrystallization are common features of subtidal cements. Commonly, high magnesium calcite replaces shell material. Pelleted carbonate mud is aggraded to a more even textured 5 to 10 micron wide microspar with aggradation beginning at the margins of the pellets (Shinn, 1969). Unit A, of the Salt River sections, is typically aggraded from a micrite to a peloidal microspar by the same process. Recementation of the fragments of deformed submarine pseudoanticlines by aragonite cement is common and can occur very quickly in certain marine environments (Shinn, 1969). Recementation of the fragmented crests of pseudoanticlines in unit A is evident (Fig. 16B). It is this process of early aragonite cementation which is believed to be the main genetic factor in the formation of submarine pseudoanticlines. The abundance of sparry calcite, as well as floating textures, in the Salt River pseudoanticlines of unit A also suggests that these structures are genetically the result of cementation (Figs. 16E, 16F).

Borings, inverted borings and encrustations are common features of submarine lithified crusts and

submarine pseudoanticlines (Lindstrom, 1963; Shinn, 1969; Assereto and Kendall, 1977). Extensive burrowing of subtidal sediment, in association with pseudoanticlines, may be useful for distinguishing submarine pseudoanticlines from other pseudoanticline types (Shinn, 1969). "Destruction of sedimentary structures by burrowing organisms can be relatively more effective in areas of slow sedimentation, and it follows that sediments deposited slowly enough to become submarine cemented are extensively burrowed and generally lack good sedimentary lamination" (Shinn, 1969, p. 123). Extensively burrowed lithologies are to be expected in submarine pseudoanticline structures. In contrast, the pseudoanticlines occurring in unit A (section H) are not burrowed, but are faintly laminated.

CRESTAL DEFORMATION AND SEDIMENTARY FILL OF SUBMARINE PSEUDOANTICLINES

Axial fractures in submarine pseudoanticlines are, millimeter to meter sized, severely fragmented, and overthrust zones are characterized by surfaces of cemented rock flour (Shinn, 1969). Lindstrom (1963) also reported fractured and thrust pseudoanticline crests from Ordovician examples in Sweden. Axial fractures commonly occur in the Salt River pseudoanticlines from unit A (Fig. 16C).

Fracture fill cements are not common in submarine pseudoanticlines, but fibrous aragonite hemispheroids

and aragonitic microcrystalline limestone do occur (Assereto and Kendall, 1977). However, submarine pseudoanticline fractures can also be filled with various internal sediment. "Cemented internal sediment is one of the most common and significant features of submarine-cemented rocks in the Persian Gulf" (Shinn, 1969, p. 137). Highly fractured and buckled axial crests that contain cemented internal sediment commonly occur in the small pseudoanticlines found in unit A along the Salt River (Figs. 16B, 16D).

SUBMARINE PSEUDOANTICLINE DESIGNATION

Based on comparisons with those described in literature the centimeter sized pseudoanticlines that occur in unit A are herein considered to be submarine pseudoanticlines.

PSEUDOANTICLINES OF UNITS B AND C

The pseudoanticlines in units B and C are comparable to tepee structures reported in literature. Typically, the crests of most tepees connect to form large, four to five sided megapolygons that generally have triradiate polygonal sutures in plan view. Most of the tepees contained in the coastal salinas of Southern Australia have a non-orthogonal pattern; however, orthogonal patterns do occur (Warren, 1982). Polygons occur in a regular pattern (Davis, 1970; Evamy, 1973; Ferguson et al., 1982; Warren, 1982) or an irregular pattern (Burri, 1973). Newall et al. (1953) suggested that tepees are linear features lacking preferred

orientation. Of all the patterns the polygonal pattern is generally the more common pattern. In Holocene examples from Southern Australia, megapolygons occur associated with large, domed (convex upward) oval centers of lithified Holocene sediment (Ferguson et al., 1982; Warren, 1982). The size of the megapolygons decreases towards the center of the lithified zones, as does the thickness of the lithified crusts in which the tepees are formed (Ferguson et al., 1982; Warren, 1982). In the center of the polygons smaller polygonal systems (1 to 5 m) are generally superimposed on the large tepee bound polygons. The thickness of the lithified crust in the center of the polygons and near the edge of the over-riding polygonal margins (tepee structures) is 10 and 40 cm, respectively (Ferguson et al., 1982; Warren, 1982). Comparable Recent and ancient tepee structures were reported by Assereto and Kendall (1977).

The polygonal pattern, in cross-section, represents the upwarping margins of "saucers". The term "saucers", as used in this study, refers to the inter-tepee areas. Together, the tepees and related inter-tepee areas form the characteristic polygonal pattern seen in Recent and ancient examples. Geometrically, saucers have flat to slightly arched centers and curve upwards to 45° at their margins (Assereto and Kendall, 1977). It is the upturned margins of the saucer that form the tepee structures. Generally, the slopes of the tepee (saucer margins) dip at less than 50° , but, on higher portions of the structure, dips can be

vertical or slightly overturned (Smith, 1974).

Generally the saucers range in size from one to a few tens of meters in height and up to 10 m in width. Variations in the dimensions of the saucers are generally related to the thickness of the deformed sediment (Assereto and Kendall, 1977). Deformed beds in the saucers range from 20 to 40 cm thick in those saucers with widths < 3 m to 4 m thick in the larger 10 m wide saucers.

In cross-section, tepees can form simple structures a few cm high and wide or extremely complex structures up to a few tens of m in height and width. Tepees in the Upper Devonian of the Guadalupe Mountains reached a height of 9.1 m and a width of 18.3 m (Smith, 1974). In a study of the Triassic Calcare Rosso of Lombardy, Italy, the ratio between the diameter of the base and the height of a tepee varies from 4:1 to 5:7 (Assereto and Kendall, 1977). Relief of individual beds rarely exceeds a few feet (Smith, 1974), with the base of a tepee generally consisting of lower beds that are either flat or gently bent upwards to form long narrow folds (Assereto and Kendall, 1971, 1977; Smith, 1974; Burri et al., 1973). The upper beds of a tepee are more extensively deformed than the lower beds. In Holocene examples lithified crusts actually override one another. The crusts are commonly brecciated, extensively fractured and/or faulted (Davis, 1970; Evamy 1973; Ferguson et al., 1982; Warren, 1982). Similar observations were reported from ancient examples (Assereto and Kendall, 1971, 1977; Smith,

1974; Burri et al., 1974), as well as from the Salt River pseudoanticlines.

The pseudoanticlines in units B and C are not visible in plan view, however, the typical tepee shape and the formation of saucers is evident (Figs. 16G, 19A). The saucers formed in the Salt River units are 0.2 to 10 m wide, therefore are similar in size to tepee saucers reported by Assereto and Kendall (1977). However, the ratio between the diameter of the base and the height of the tepee is below that designated by Assereto and Kendall (1977). The moderately sized pseudoanticlines might represent smaller tepees superimposed on larger polygons, which are represented by the large antiform structures PA4, PA6, PA7 and PA8 (Appendix 3).

The complexity of the tepee, expressed in terms of increased brecciation, faulting, fracturing, and number of erosion surfaces, is proportional to the size of the tepee. The larger the tepee the more complex its nature. "From the interrelations of component beds, it is clear that the larger examples [referring to tepees] had a long and varied history, and that phases of growth of the structure and accompanying brecciation were separated by phases of further sediment deposition and cementation and by periods of contemporaneous erosion" (Smith, 1974, p. 64). The higher pseudoanticlines can consist of a number of small tepees, separated by erosion surfaces, that are developed one above another (Smith, 1974). Tepees rarely develop on top of an

eroded tepee crest; however, tepees with the same orientation can develop on crests that are not eroded (Smith, 1974). In one example, the end result is a cedar-tree-like structure that is more than 9.1 m high (Smith, 1974). Similar structures in the Lower Jurassic of Morocco, Africa (Eastern Atlas) comprise stacked layers of tepees that alternate with non-deformed beds and reach thicknesses of several tens of meters (Burri et al., 1973). Truncated tepee crests, that result in erosional surfaces that are overlain by horizontal strata, were reported by Assereto and Kendall (1977).

There is no modern analog of "stacked tepee structures", nor is there a good Holocene example of truncated tepee crests and subsequent deposition of overlying strata. It is certainly possible that these Recent structures did not have enough time to develop to the degree of complexity seen in tepees of the ancient record.

Some of the Salt River antiforms are composite structures similar to composite tepees reported from literature (Fig. 18). The complexity of the tepee is proportional to the size of the tepee in the Salt River examples. Brecciation, fracturing and the number of erosion surfaces become more abundant in the larger pseudoanticlines, with phases of growth separated by further phases of sedimentation.

TEPEE LITHOLOGIES

Tepees occur in a variety of carbonate lithologies, but lithologies generally reflect intratidal/supratidal sedimentation along a coastline or in a restricted coastal salina where very shallow water conditions and periodical wetting and drying of the sediment can occur. Tepees are common in marine vadose fenestral and pisolitic limestones and/or dolomites that range in age from Silurian to Holocene (Assereto and Kendall, 1977).

The most commonly occurring feature in lithofacies associated with tepee structures is fenestral porosity. In Holocene examples, there is good evidence to indicate that fenestrae units are cemented with an early aragonitic cement (Ferguson *et al.*, 1982; Davis, 1970; Evamy, 1973; Warren, 1982). Fenestrae units are also reported from ancient examples. Ancient tepees occur in boxwork boundstone in the Permian Capitan Reef Formation (Warren, 1982); in fenestral carbonate rocks, fine doloarenites and dolorudites, of the eastern Guadalupe Mountains (Smith, 1974); in laminar horizons of highly altered fenestral intramicrites and in indurated crusts of supratidal, intraclastic, partially dolomitized fenestral pelmicrites (Assereto and Kendall, 1971, 1977).

There are many different types of fenestrae associated with tepees in the Triassic Calcare Rosso of Lombardia, Italy (Assereto and Kendall, 1977). The two

types of fenestrae are, finely laminated fenestrae that are arranged horizontally to form subparallel, closely spaced voids and fenestrae consisting of larger, more irregularly shaped cavities randomly oriented and distributed throughout the rock (Assereto and Kendall, 1977). Larger fenestrae can form elongate 2 to 5 mm wide cavities (Ferguson et al., 1982) and voids reaching hundreds of cubic cm in volume (Warren, 1982). The intertidal/supratidal, highly dolomitized, highly recrystallized and fenestral lithologies in the Salt River pseudoanticlines are similar to tepee related lithologies described in the literature. Large vugs and high porosities also commonly occur (Fig. 19B).

Speleothem deposits which occur in Holocene tepees in the Fisherman Bay area are of 3 types, namely, pisolites, flow aragonite and aragonitic pool deposits (Ferguson et al., 1982).

Intertidal and supratidal pisolites are generally associated with Holocene tepees and cave deposits (Shinn, 1973; Purser and Loreau, 1973; Scholle and Kinsman, 1974; Esteban, 1976, Ferguson et al., 1982). However, documentation of pisolites associated with ancient tepees is rare.

Of the flow aragonite deposits there are two distinct cavity deposits reported from areas where Holocene tepees are forming; namely, those that form on the roof and on the floor of a cavity. There is a marked

difference between the roof and floor coatings of these cavities (Ferguson *et al.*, 1982).

Cavity deposits can form a high proportion of the total volume of tepees. Up to 80% of the carbonate studied at Fishermans Bay, South Australia is a "cavity deposit" (Ferguson *et al.*, 1982). Generally, the cavities are connected, and can form complex intricate conduits throughout a Holocene lithified crustal.

Like the Holocene counterparts, ancient examples of cavity fills related to tepee structures can also be divided into floor and roof deposits. Paleospeleothems and cements of the Triassic Calcare Rosso of Lombardo, Italy that were originally comprised of aragonite, later inverted to calcite, are represented by seven, commonly occurring ancient cement textures; namely, festooned cellular crusts, giant paleo-aragonitic rays (raggioni), microscopic paleo-aragonitic crusts, paleo-aragonitic hemispheres, coconut-meat calcite, dolomitic microstalactites and gnocchi (Folk and Assereto, 1974; Assereto and Kendall, 1977; Assereto and Folk, 1980).

Exotic fibrous cellular aragonite cements are common in tepees. However, such cements are lacking in units B and C of the Salt River pseudoanticlines. The absence of this important tepee diagnostic property may be elucidated by the fact that extensive diagenesis, shown to exist in the Salt River structures, has completely destroyed any trace of aragonite, exotic

cement forms and post aragonite textures. The diagenetic history of the structures is complex and complete.

INTERBEDDED LITHOLOGIES OF TEPEES

Interbedded lithologies commonly associated with tepee structures range from lithologies representative of subtidal to supratidal environments to terra rossa layers (Assereto and Kendall; 1971, 1977). Commonly occurring in association with tepees are stromatolites and related algal facies (Ferguson et al., 1982; Warren, 1982; Davies, 1970; Shinn, 1967; Burri, 1974; Smith, 1974, Assereto and Kendall 1977), bioturbated fossiliferous micrite (Assereto and Kendall, 1977) and aragonitic crusts (Evamy, 1973; Purser and Loreau, 1973). In the Salt River strata bioturbated facies are evident, but algae only occur in small amounts.

Evaporites (mainly gypsum) are commonly associated with tepee structures. Aeolian gypsarenite and shallow water laminated gypsum deposits occur in coastal salinas of South Australia in association with tepees (Warren, 1982). Gypsum discs, gypsum and halite crusts generally occur only on the surface of polygons from the Fisherman Bay area in South Australia (Ferguson et al., 1982).

Aeolian dune sands and ostracod-pellet wackestones occur on the margins of coastal salinas (Warren, 1982). In the Fisherman Bay area eolian Pleistocene dune sands were also reported (Ferguson et al., 1982).

The rocks associated with pseudoanticlines should be located as a sequence of horizons on the margins of a carbonate platform (Assereto and Kendall, 1977). In this study, the lithologies are representative of subtidal to supratidal conditions that existed on a shallow carbonate platform. The platform occurred on the northern edge of the Devonian La Crete Basin. Assereto and Kendall (1977) suggested that tepees form in peritidal sediments that are more or less altered by schizohaline vadose diagenesis. The Salt River pseudoanticlines formed in an intertidal to supratidal environment where mixing of marine and meteoric waters resulted in three phases of dolomitization and increasing amounts of fresh water caused subsequent phases of dedolomitization.

CRESTAL DEFORMATION OF TEPEES

Tepees commonly contain crestal axial fractures and/or axial faults. These disruptions may have formed by tensional separation of adjacent polygonal plates or by compressional thrusting or buckling of semi-lithified crusts. Since the genesis of these axial deformations and openings are speculative, only the physical attributes of these breaks can be discussed.

Axial fractures of Holocene tepee structures are common (Davies, 1970; Shinn, 1969; Assereto and Kendall, 1977; Ferguson et al., 1982; Warren, 1982). The axial fractures generally occur as two types (Assereto and

Kendall, 1971, 1977). In simple tepee structures the fractures occur parallel to bedding. In complex tepee forms, the axial breaks are more complex, displaying sub-vertical fractures that cut across the fill of earlier cracks (Assereto and Kendall, 1977). The subparallel cracks are generally several meters in length and up to 80 cm in width (Assereto and Kendall, 1977). Subvertical axial fractures also occur in tepees from the ancient record (Smith, 1974). Having studied several hundred tepee structures from Permian Shelf carbonate rocks of the Guadalupe Mountains, New Mexico, Smith (1974) found very few tepees displayed strictly vertical, planar axial fractures. If the margins of the crestal break are sharp and mirror images of one another, then the break is classified as an axial fracture. If opposite borders of the crestal break lack symmetry, then the opening is either a result of faulting and/or dissolution (Assereto and Kendall, 1977). Both axial fractures and faults occur in the Salt River pseudoanticlines.

CRESTAL FILL OF TEPEES

Subvertical fractures commonly have an asymmetric fill whereas vertical fractures exhibit a more symmetrical fill (Assereto and Kendall, 1977). Fissures occurring in Holocene tepees from the Fisherman Bay area, South Australia are lined with radial aragonite, lacy white carbonate and contain pisolites (Ferguson et

al., 1982). Lacy white carbonate lined fractures also occur in Ancient Triassic tepee structures (Assereto and Kendall, 1982). In both ancient and Recent examples, the cement fill and/or sedimentary fill that occurs in the crestal breaks may be vertically laminated (Ferguson et al., 1982).

Internal sedimentary fill in openings, axial fractures or axial faults can range from various cement types to actual non-lithified sediment that was washed into the original void space. This sediment can consist of both allochemical or terrigenous grains (Assereto and Kendall, 1977). The most common type of fill is a red noncalcareous mudstone that is variably mixed with micrite and pelletoidal micrite or terrigenous siltstone and sandstone (Assereto and Kendall, 1977). Laminar, fibrous micrite and terrigenous sand and silt filled tepees of the Gaudalupe Mountains (Smith, 1974). Carbonate silt, derived either from direct precipitation in the void areas, or from erosion of the carbonate lined walls is another possible fill.

The bioclastic debris and sediment that is washed into the fractures of the tepee reflects the age and ecosystem that persisted at the surface or in the cavities of the tepee (Assereto and Kendall, 1977). Generally, this sediment is washed in from the surface penecontemporaneous with, or shortly after the openings of the cracks. A Holocene analog to this is given by

tepees forming in the Fisherman Bay area, South Australia (Ferguson et al., 1982). There, small chips of eroded pavement (lithified crust) not only occur on the surface of the saucers, but also accumulate with clay in fissures forming along the polygonal margins (tepee crests).

The internal sediment that is washed into the void spaces of the tepees can exhibit faint parallel lamination and fining upward sequences that are suggestive of deposition under waning current conditions. Cut and fill structures, ripple drift lamination and coarse cross-bedding are suggestive of stronger currents that may have flowed through the cavities, fractures and other void spaces inherent in the tepee. Crestal fill in the Salt River pseudoanticlines is evident (PA6 especially) but the source of the fill with respect to provenance and age is not discernible.

In both submarine and tepee pseudoanticline forms the fracture fill sediment is penecontemporaneous with the deformational event. The age of the overlying beds are younger and differences in age are not detectable, paleontologically in the tepee form (Assereto and Kendall, 1977). The above statements are also true of the pseudoanticlines formed along the Salt River.

TEPEE DESIGNATION

The above evidence suggests that the Salt River pseudoanticlines are extremely similar to tepee structures. Referring to the definition of tepees, as given by Assereto and Kendall (1977), it has been shown that the Salt River structures meet the criteria needed to classify them as tepees. It has been shown that the antiform structures have formed by penecontemporaneous expansion on the margins of a near marine environment. Although the base of the larger tepees was not observed in the Salt River area it is unlikely that the underlying gypsum caused expansion of the Salt River Strata. Expansion was probably due to a fresh water lens that existed beneath the strata containing tepees. This lens would have been more buoyant than the surrounding seawater, thus doming the area and producing the pseudoanticlines. The composite nature of some of the tepees suggests that there was more than one deformational event. Repetitive deformation might reflect seasonal changes in hydrology of the fresh water lens. A domed nature of the Salt River strata is also evident. Whether or not this domed nature represents the original doming of the strata due to a buoyant underlying fresh water lens, or later tectonic movements, is not certain.

Evidence of fresh water diagenesis and subaerial exposure suggests that the tepees along the Salt River

had formed during the Middle Devonian in environments similar to tepee environments described in South Australia by Ferguson et al. (1982). Subaerial exposure was in response to repeated regressions of the Middle Devonian seas.

Based on the previous statements, the Salt River antiforms in units B and C are herein thought to be pseudoanticlines of the tepee type. Because of their complex brecciated nature, they possibly represent the mature tepee form.

VI. COLLAPSE OF THE SALT RIVER STRATA

A. COLLAPSE BRECCIAS

PHYSICAL DESCRIPTION OF THE BRECCIAS

Some of the breccias that occur along the Salt River represent a collapsing event that occurred after the formation of the pseudoanticlines. Breccias which are genetically not the result of pseudoanticline formation occur, in sections D, G, I, K and P. Sections I and P contain breccias which are most obviously of a collapse origin.

Section I is a clast supported particulate rubble packbreccia that is 20 m wide and 6 m high (Figs. 20A, 20B). Both west and east boundaries of the breccia are with non-brecciated strata. Both boundaries are well defined, irregular surfaces (Fig. 20B).

Blocks in the breccia are up to 2.75 by 1.15 m in size. The larger blocks have a modal size of 0.3 m (maximum dimension). Most blocks are angular and at the boundaries of the breccia the blocks are elongated parallel to the dip of the strata entering the breccia. Towards the center of the breccia the block orientation becomes more random. There are a few large conformably bedded blocks in the breccia. A till layer, which is 0.7 m thick, occurs above the breccia.

Both block and matrix lithologies are comprised of units A, B and C. The matrix contains a high amount of black shale that weathers to a pale green to rust color. Pockets

of shale occur in breccia of sections D and P. Brachiopods and bivalves, identical to those located in units A through C, occur in both the clasts and the matrix.

There are two lithologies which occur in the breccias but not in the measured sections along the Salt River. One lithology is the dark boudinaged unit previously defined as the Keg River marker. It occurs in breccia in sections D and I. The other lithology is represented by the largest block in section I (Fig. 20B). This block is composed of finely crystalline, oncolitic dolostone. Dirty, anhedral to euhedral 31 micron diameter crystals of dolomite occur in a sucrosic mosaic. Relict calcite occurs up to 4% of the rock. Ghost textures of brachiopods, gastropods and echinoderms occur as trace amounts up to 3% of the rock. Oncolites form up to 15% of the lithology and can attain lengths of 4.5 cm and radii of 1.6 cm. Diagenesis was extensive and the origin of the nucleus can not be determined. The oncolites are 90% dolomite and 10% calcite. The dolomite rhombs increase in size and become more euhedral towards the outer edges of the oncolites, whereas the calcite is concentrated in poorly defined layers closer to the outer edge. The poorly defined layers are probably algal related and form 4 or 5 concentric rings around the nucleus. This unit is lithologically similar to the basal part of the Upper Keg River Member (unit D).

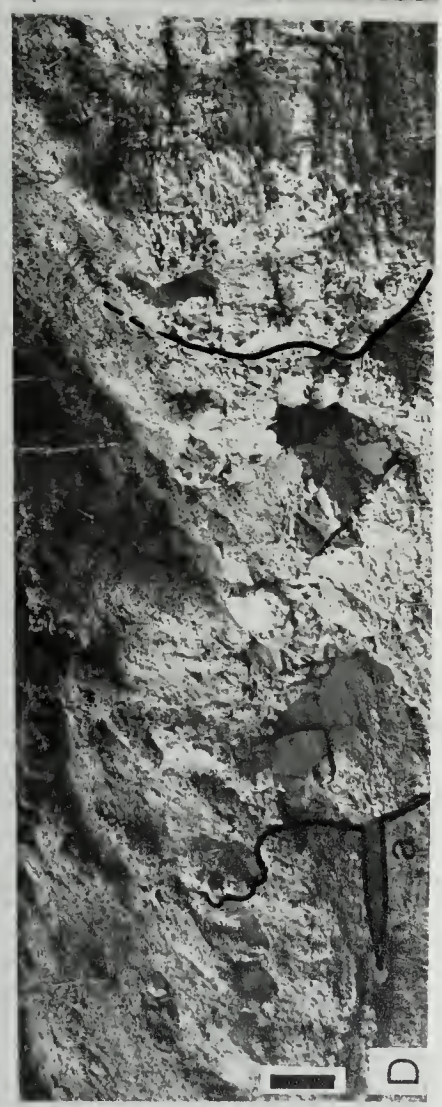
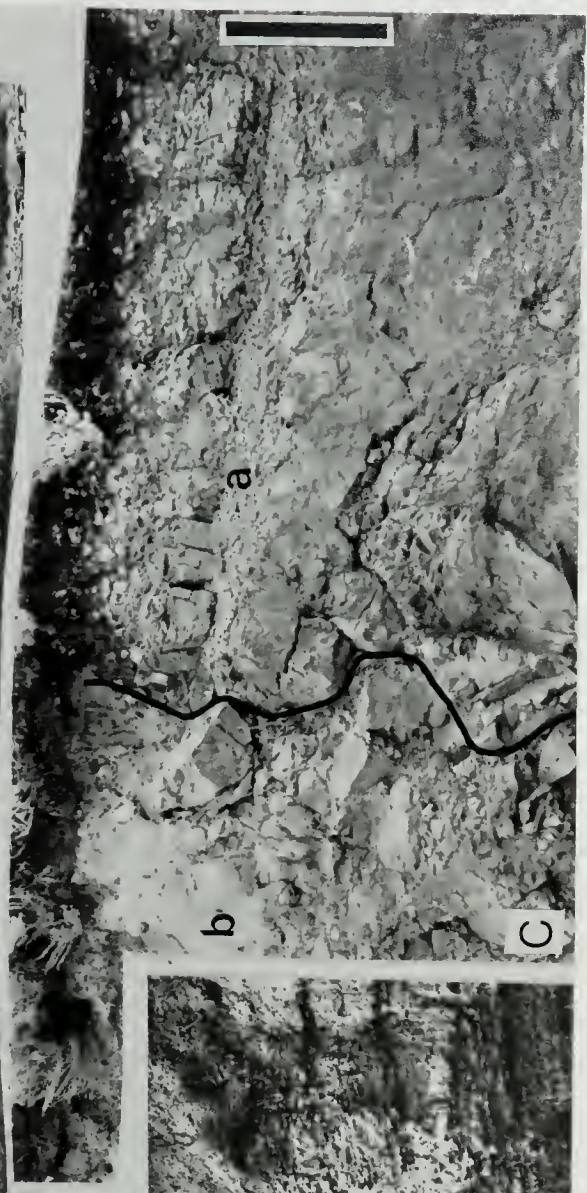
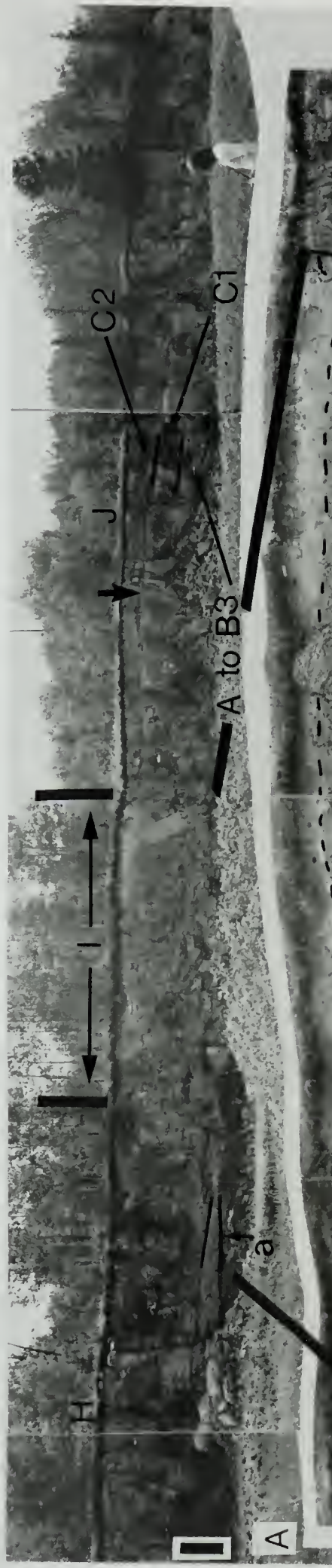
FIGURE 20

20A) Field photograph of sections H, I and J showing the laterally restricted extent of section I (collapse breccia) and the bedded strata of sections H and J on either side of section I. Units A through C2 are (black heavy arrow) pointing to the recessive biomicrite marker. Unit B1 pinches out against the eastern edge of section I. Scale bar equals 1.5 m.

20B) Field photograph of section I. The solid black lines define the lateral extent of the breccia. The small black arrows define bedding in unit A, as unit A is traced from bedded strata outside the breccia into the breccia, where a pseudoanticline structure is defined. Typically, collapse breccia occurs over the top of the pseudoanticlines. Large blocks of bedded strata occur in the breccia (a). A 2.75 by 1.15 m oncolitic dolostone block (b) represents unit D. Both unit D and the keg river marker represent younger strata that underwent collapse. The clasts become relatively younger towards the west. The biomicrite marker (c) abruptly ends at the western contact. Overburden was deposited over the breccia and over the west end of section J (dashed line).

20C) Field photograph of western contact of section I (solid dark line). The recessive biomicrite marker (a) abruptly ends at the contact. The overburden (b) does not occur in section J, but is abruptly terminated at the contact. Scale bar equals 0.5 m.

20D) Field photograph of a collapse breccia (section D). The breccia is laterally restricted. The dark solid lines represent the contact between the breccia and unit C2. A fissure extends into unit C2 on the western edge (a). Scale Bar equals 1 m.



clasts become younger

MICROSCOPIC FEATURES

Numerous stylolites, geopetal textures and microfaults occur throughout the breccia. Elsewhere in the Salt River sections these features are neither as evident, nor as abundant.

Stylolites with amplitudes up to 3 cm are slightly yellow to medium brown and highly concentrated with an organic residue. A loss of calcite at the stylolites is supported by the abrupt truncation of fossils and clasts on either side. Also, near-matching portions of the allochems are observed on either side of the stylolite. However, an aggraded calcite occurs between organic residue of the stylolite and appears to have split the stylolite in half. The calcite is anhedral and dusty (not dirty like the aggraded calcite of the matrix). As the pocket of calcite grew the crystal size increased into the center of the pocket. Many of the numerous calcite lined fractures which occur in the breccia possibly originated as small pockets of calcite along the stylolite. Commonly, stylolites appear to change laterally into the calcite lined fractures. On the edge of the fractures stylolites appear to be forcing pieces of the host material away from the host rock. As precipitation or aggradation of the calcite continued, that portion of the host was completely "plucked away" from the edge of the fracture. It then became a floating clast, completely surrounded by calcite. In agreement with Watts (1978, 1979), displacive calcite appears to be a viable

force in many of the thin sections studied from section I. It is realized that the third dimension is not observed in a thin section, but the plucking apart of the host material on the edges of fractures and along stylolites occurs repetitively. There has always been questions as to where the dissolved calcite at a stylolitic surface goes to. In the case of the Salt River collapse breccias it would appear that the calcite dissolved at the stylolitic contact is reprecipitated elsewhere along the same stylolite. In the Salt River collapse breccias the calcite appears to display "displacive forces".

B. RELATIONSHIP BETWEEN COLLAPSE BRECCIAS AND PSEUDOANTICLINES

The relationship between the collapse breccia and the pseudoanticlines is best seen in sections E, G and I. There is a relative degree of collapse that occurs among these three sections. The collapsing phase is superimposed on top of the pseudoanticline deforming event. In section E there was little to no collapse. Vertical fissures filled with breccia are always concentrated above the pseudoanticlines (Fig. 16G, 19A). In section G, collapse is more obvious, consequently the strata are much more brecciated with indistinct fissures occurring above the larger pseudoanticlines (Fig. 19C). Above PA4 the distinct shape of the tepee can be followed. However, on the south side of the tepee unit C2 becomes highly brecciated (Fig. 18, Appendix

3). In section I collapse is at its maximum. But, upon closer examination, units A and B of section H bend upward as they enter section I (Fig. 20B). Other fabrics, such as a row of calcite vugs also extend upward into section I. The bending upward of the strata is definitely not expected to occur in a breccia thought to be of a collapse origin. However, the breccia is obviously a collapsed structure due to its limited lateral extent and the fact that younger horizons (Keg River marker and Upper Keg River Member) are collapsed down, laterally adjacent to older strata represented by unit A through C. Tracing unit A from section H into the breccia (section I), it is obvious that this structure forms a relict pseudoanticline. This pseudoanticline is not as physically evident as pseudoanticlines in sections E & G because the collapse is much more extensive in section I. From sections E to G, to section I the pseudoanticline structures become less apparent because the degree of collapse (the degree of brecciation) increases. As with PA4 (section G, Fig. 18), the collapse in section I increases away from the pseudoanticline. Towards the west of section I the collapsed strata become younger within the breccia (Fig. 20B).

C. COLLAPSE ORIGIN

The breccia of section I and portions of sections G and E are of a collapse origin. Dissolution of the tepees and saucers is evident from the outcrops along the Salt River.

However, is collapse due to dissolution of the above structures, or dissolution of the underlying evaporite, or both? It is highly possible that the preexisting pseudoanticlines acted as sites of preferential dissolution. But, the main collapsed areas are adjacent to the tepees, in the saucers. Perhaps the dissolving fluids were channelled between the tepees, into the saucers, where the main dissolution occurred.

Other problems occurring with these later formed breccias are, what was the origin of the dissolving fluids and what was the age of the collapse? Were the fluids dissolving the Keg River and Chinchaga strata beneath a subaerially exposed paleosurface during the Devonian or were the dissolving fluids a result of a more Recent groundwater phenomena?

Corrigan (1975) stated that in the La Crete basin there were numerous sea level drops in the Upper Keg River Member that he thought caused surface exposure of the strata and subsequent collapse by solution in the vadose zone (chapter 2). It stands to reason that these sea level drops would be more pronounced on the edge of the La Crete Basin than in the center where Corrigan (1975) documented them. Because the Salt River area is believed to represent the northernmost edge of the La Crete Basin sea level drops should be relatively easy to detect. Corrigan (1975) substantiated the sea level drops and horizons of subaerial exposure by documenting synchronous collapse breccias,

oncolite horizons and calcrete horizons. The youngest rock in the breccia is the oncolitic dolostone of the Upper Keg River Member, thus implying that there may have been a subaerial surface above the basal oncolitic dolostone of the Upper Keg River Member. Such a surface was documented by Corrigan (1975) in the La Crete Basin.

Clasts in the breccia in section I are extremely fractured. The fractures would obviously be filled with the next youngest sediment deposited over the brecciated strata. The fractures are filled with fragmental pieces of rock identical to the Lower and Upper Members of the Keg River Formation. No younger strata are observed. Some of the fossils occurring in the fractures and matrix are two holed crinoidal ossicles (Gasterocoma) and Atrypa. Some of the fossils were whole (not fractured) and did not occur in a fragmented host lithology suggesting that they were a sediment and not a rock when they were incorporated into the fractures. Also, shale lined fissures in sections E and G suggest that water was percolating down from above the pseudoanticlines and washing down a brown to black silty shale. This water which may have provided the "drusy cement" in unit C2, may have been associated with a subaerially exposed surface higher in the section.

Tsui (1982) suggested that the breccia of section I is paleokarst and formed prior to the last glaciation (Wisconsin Laurentide ice sheet). The age of collapse of the strata along the Salt River occurred after deposition of the

basal deposits of the Upper Keg River Member because both the Keg River marker and the oncolitic dolostone of the Upper Keg River Member occur in the breccia. Based on the above evidence the collapse probably occurred prior to the end of the Devonian Period with the breccia formed as a result of subaerial exposure at the end of deposition of the oncolitic dolostone or shortly (geologically speaking) thereafter.

However, one piece of evidence which complicates matters is that there is 0.7 m of overburden above the Devonian strata in section I (Fig. 20B). In the adjacent section J there is no overburden. To the east the overburden horizon can be traced a few meters out of section I and into section H, where it quickly thins. The presence of this capping overburden unit suggests that collapse occurred after the till was deposited (i.e. after the last glaciation). But, because the till is not confined solely to the breccia (it also occurs on the western edge of section H) this 0.7 meter horizon may have been the result of selective erosion by the advancing glaciers. The breccias in this section of the Salt River strata may have been more susceptible to erosion by the glaciers. Afterwards, till could have easily been deposited into the depression caused by the glaciers.

Section I, and possibly other breccias along the Salt River (Fig. 20D) that are not the result of pseudoanticline formation, are representative of paleokarst. Due to the

restricted lateral extent, shape, chaotic brecciated nature and similar stratigraphic position of suberially exposed surfaces and collapse breccias in the La Crete Basin (Chapter 2) the breccias probably represent paleodolines that formed shortly after the basal deposition of the Upper Keg River Member. Those strata which occurred above were removed by later glaciation.

VII. STRATIGRAPHY OF THE FORT VERMILION/SLAVE POINT FORMATIONS OF PEACE POINT, WOOD BUFFALO NATIONAL PARK

A. LITHOFACIES

INTRODUCTION

In the study area (Fig. 2), the Fort Vermilion and Slave Point Formation comprise ten lithofacies; namely, (1) the bedded evaporite lithofacies, (2) the brown limestone/intermixed dolostone lithofacies, (3) the gypsum mould breccia lithofacies, (4) the blocky breccia lithofacies, (5) the shaly breccia lithofacies, (6) the gypsum breccia lithofacies, (7) the breccia I lithofacies, (8) the breccia II lithofacies, (9) the breccia III lithofacies, and (10) the sandstone breccia lithofacies. Although the ten lithofacies are distinct field units they are in part gradational into one another. Some of the lithofacies represent a gradational spectrum in which diagenetic changes and the degree of brecciation becomes more complex. For example, lithofacies 2a represents the original unaltered unit in the study area. As diagenesis began the limestone of lithofacies 2a was gradationally recrystallized to form lithofacies 2b. Lithofacies 2c represents the next stage along the spectrum where the recrystallized limestone of 2b was partially to wholly replaced with dolomite. On the edges of the main breccia bodies in the study area lithofacies 2c develops a ropey, stretched, convoluted texture that in some areas becomes

brecciated, thereby forming lithofacies 2d. Following this gradational spectrum, lithofacies 2 changes laterally and/or vertically to lithofacies 7, which in turn is gradational into lithofacies 8. Not all of the lithofacies are gradational into one another, but for the most part a gradational aspect is evident in the field.

Panoramic photographs were taken of the cliffs of this study, from which laterally continuous drawings have been produced (Appendix 4). The drawings have been provided in the hope that the field relationships among the lithofacies can be visualized.

LITHOFACIES 1 (BEDDED EVAPORITES)

Lithofacies 1, the bedded portion of the Fort Vermilion Formation, is a recessive to prominent, white to brownish buff gypsum that contains various shades of yellow, orange, purple, grey and brown. Crystals of gypsum, only a few mm in length, commonly occur as acicular radiating and botryoidal forms. The gypsum is generally impure with medium to dark argillaceous material comprising up to 60% of some units. Generally, the argillaceous laminae are laterally discontinuous and locally surround the gypsum nodules giving a "chicken-wire" texture. Interbedded brownish buff dolomicrite rarely reaches 70% in some units. Anhydrite, which may form 60%, occurs as laminations, streaks and nodules throughout the formation. A nodular anhydritic gypsum commonly forms beneath the Fort Vermilion/Slave Point contact. Since this contact will be shown to be an angular

unconformity it is suggested that the common occurrence of anhydrite beneath the contact may be due to chemical alteration during the time of exposure which produced the unconformity. Lithofacies 1 is laminated to thickly bedded with bedding varying from massive, to ropey, to a highly contorted mosaic.

Gypsum veinlets and satin spar veins (up to 3 cm wide) are concordant or discordant. The discordant gypsum veins are numerous and suggest the gypsum was remobilized after deposition. Later episodes of gypsum intrusions, displaying "flower-like" structures, occur in gypsum blocks exposed on the river bank, beneath the Peace Point cliffs. Gypsum veins also cut across some of the breccias.

There are two marker beds which aided in measuring the Fort Vermilion sections. These markers are found in different sections and are not correlatable outside the section they occur in. One marker, which occurs in section J, is a semiprominent, 0.2 m thick, white to dark grey, multi-colored, argillaceous, anhydritic gypsum that is comprised of 6 mm long acicular gypsum crystals. Bedding ranges from distorted bedded mosaic to contorted mosaic. Section PU contains a recessive, 0.1 m thick (varies laterally), light creamy buff, planar to very slightly wavy, thickly laminated to very thinly bedded gypsum. Dolomicrite, that contains 10% nodular gypsum, comprises 30% of the section PU marker.

Localized areas of brecciation and fracturing are common, with rare faults occurring along the fractures. These features made lateral tracing of lithofacies difficult.

DISTRIBUTION AND THICKNESS

Lithofacies 1 occurs in sections PE, PH, PI, PS, PU, PX, PP, and PJ, attaining a maximum thickness of 23 m in section PU. The maximum length of lithofacies 1 is observed in sections PE, PI and PJ where lateral distances are approximately 580, 400 and 170 m, respectively. The more laterally continuous sequences occur in very broad open anticlines with flat, horizontal to subhorizontal plateau-like crests (section PE, Fig. 21A) whereas, the more laterally discontinuous outcrops occur in sections PP, PH, and PX. These smaller outcrops form tight anticlinal domes (Fig. 21C) that are occasionally overturned (section PP). Section I is a 60 m homoclinal outcrop that is fault bounded on its western side. The nature of section PS is hard to determine due to vegetation covering the outcrop.

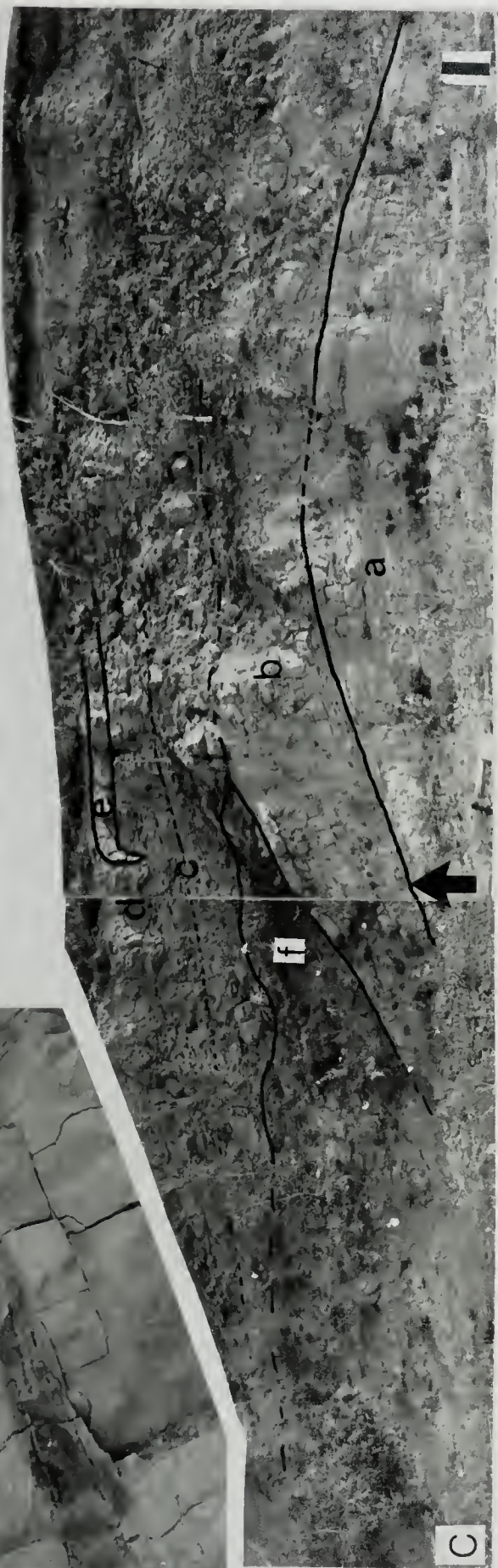
The lower contact of lithofacies 1 was not seen in the study area. However, the upper contact is erosive and an angular unconformity now separates the Fort Vermilion Formation from the Slave Point Formation. In places (section H), a recessive, medium brown to pale greenish brown, friable, calcareous claystone occurs between the two formations.

FIGURE 2 1

2 1A) Field photograph of section PE. The solid line (denoted by straight arrow) shows the lithofacies 1 (Fort Vermilion Formation, a) and the lithofacies 2 (Slave Point Formation, b and c) contact. The Slave Point Formation is divided into lithofacies 2c (massive dolostone variety, b) and lithofacies 2b (c). The curved arrow illustrates a localized collapse of lithofacies 2b above the nondeformed lithofacies 1. A nodular anhydritic gypsum unit (e) occurs just below the Fort Vermilion Formation/Slave Point Formation contact. Scale bar equal 1.5 m.

2 1B) Field photograph of lithofacies 2b (section PH) showing that the aggradation of the limestone, that produced lithofacies 2b, began along bedding planes. Penny for scale (lower left hand corner).

2 1C) Field photograph of section PH showing the Fort Vermilion Formation/Slave Point Formation contact (depicted by arrow). The domed evaporite (a) of the Fort Vermilion Formation is overlain by lithofacies 2c (massive dolostone variety, b), which is overlain by lithofacies 2b (c). Patches of lithofacies (d) occur throughout lithofacies 2b. A wedge of lithofacies 10 (f) is a recessive area comprised mainly of black carbonaceous shale with minor blocks of sandstone. Note how lithofacies 2 extends out over this wedge. The only occurrence of lithofacies 3 is in a small band at the top of section H (e). Scale bar equals 1.5 m.



Where lithofacies 1 occurs as anticlines there is very little brecciation; however, in the lows between the anticlines, there are always significant amounts of breccia that comprise lithofacies 6 and 8. Where lithofacies 1 is not developed, either lithofacies 6 (the collapsed portion of the Fort Vermilion Formation), or depending on the degree of collapse, lithofacies 8 may occur adjacent to lithofacies 1.

LITHOFACIES 2 (BROWN LIMESTONE/INTERMIXED DOLOSTONE)

Lithofacies 2 represents the bedded portions of the Slave Point Formation. Where distinguishable, the lithofacies can be divided into 4 subdivisions, namely: 2a, 2b, 2c and 2d.

LITHOFACIES 2a (MASSIVE BROWN LIMESTONE)

Lithofacies 2a includes the limestone which is the least diagenetically altered rock in the Slave Point Formation. It has not been extensively recrystallized or replaced with dolomite.

Lithofacies 2a is a semiprominent to prominent, light to dark brown, very finely to finely crystalline, limestone. Interbedded/interlaminated, argillaceous, dark brown laminae generally comprise 40% of the unit. Bedding is irregular, slightly wavy to wavy, and ranges from 1 to 10 cm, but averages about 2 cm.

Lithofacies 2a is the only lithofacies in which sedimentary structures can still be seen. Sedimentary structures (from section PE) include, mudcracks, small

scale, low angle, trough cross-stratification and semi-hemispheroidal mounds. The mounds are less than 0.5 m diameter, lack internal structure, but may be representative of stromatolites.

LITHOFACIES 2b (RECRYSTALLIZED, VUGGY LIMESTONE)

Lithofacies 2b is a recessive to prominent (generally prominent), reddish, medium brown to buff, very coarsely crystalline limestone. Beds are wavy to planar and laterally discontinuous. The wavy bedding is due to a diagenetic displacement of original bedding by extensive calcite aggradation. Calcite aggradation proceeded along bedding planes, separating the limestone lithofacies as the calcite crystals grew (Fig. 21B). In some areas (section PH) the recrystallization is so extensive that floating textures were formed with brown limestone completely surrounded by an extensively aggraded calcite (Fig. 22A). In places, vugs which formed along bedding planes completely obliterated preexisting bedding. Bedding ranges from 2 to 40 cm, but averages about 5 cm.

Extreme vugginess is characteristic of lithofacies 2b. The vugs are lined with rosette, botrydial, scalenohedral and cubic calcite forms (Figs. 22D, 22F). The vuggy texture of lithofacies 2b increases towards the base in some exposures (section PE), with the top of the exposure being less recrystallized and lithologically more similar to lithofacies 2a.

FIGURE 22

22A) Field photograph of lithofacies 2b showing that aggradation that began along bedding planes can be quite extensive. Limestone clasts are "floating" in an aggraded coarser calcite (a). Small vugs that had formed along bedding (b) can coalesce to form larger vugs (c). Scale bar equals 10 cm.

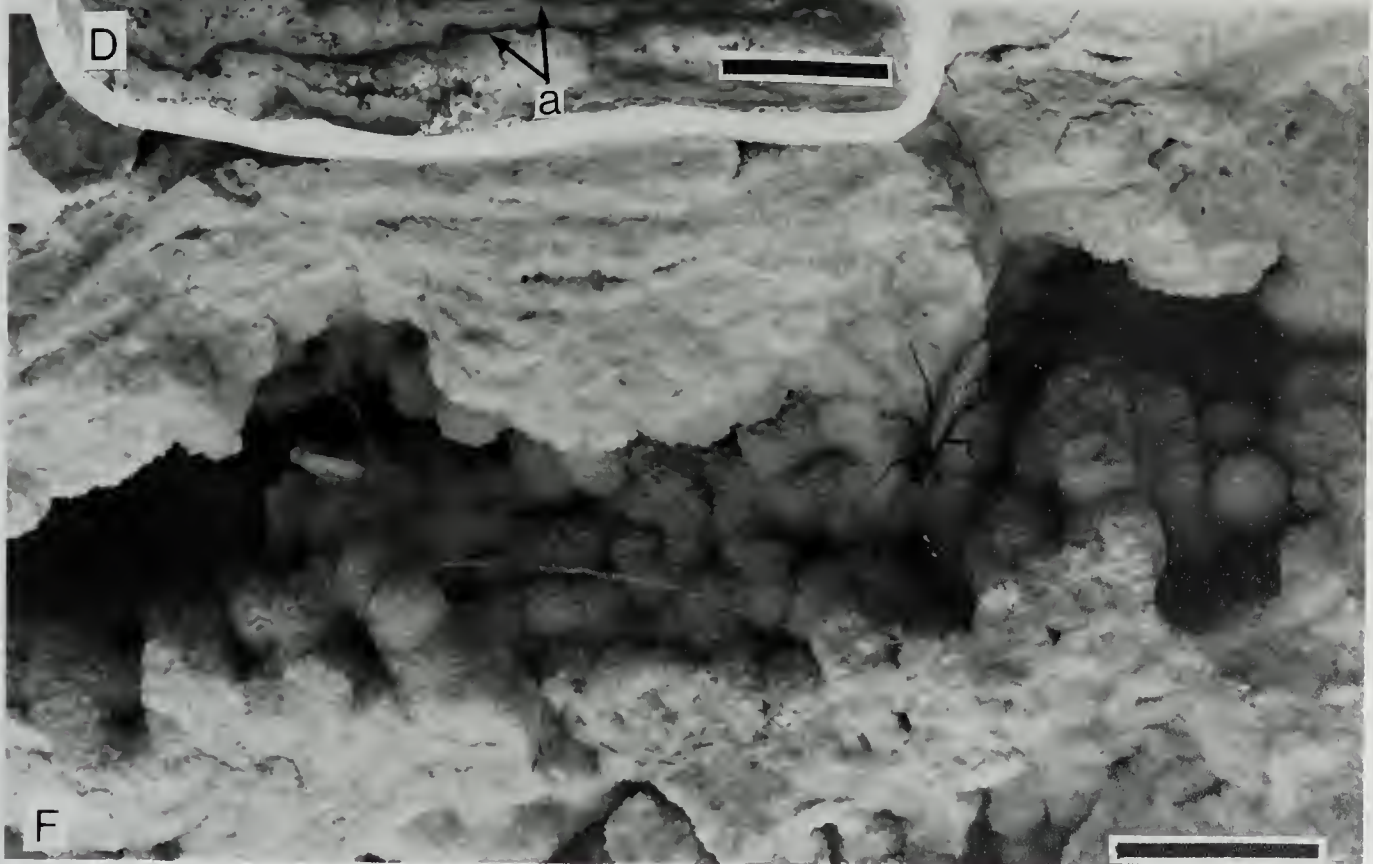
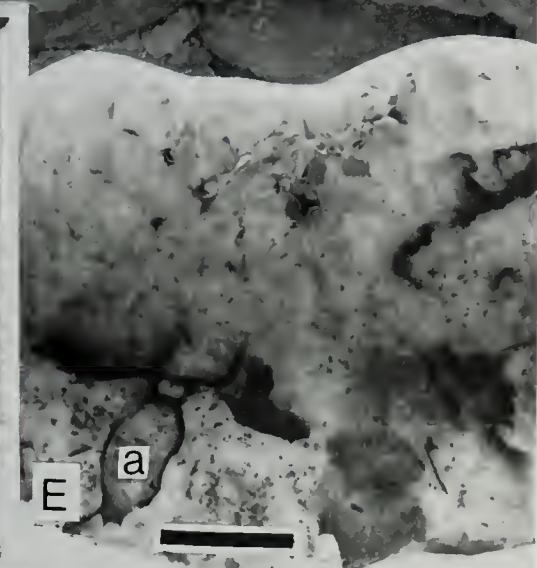
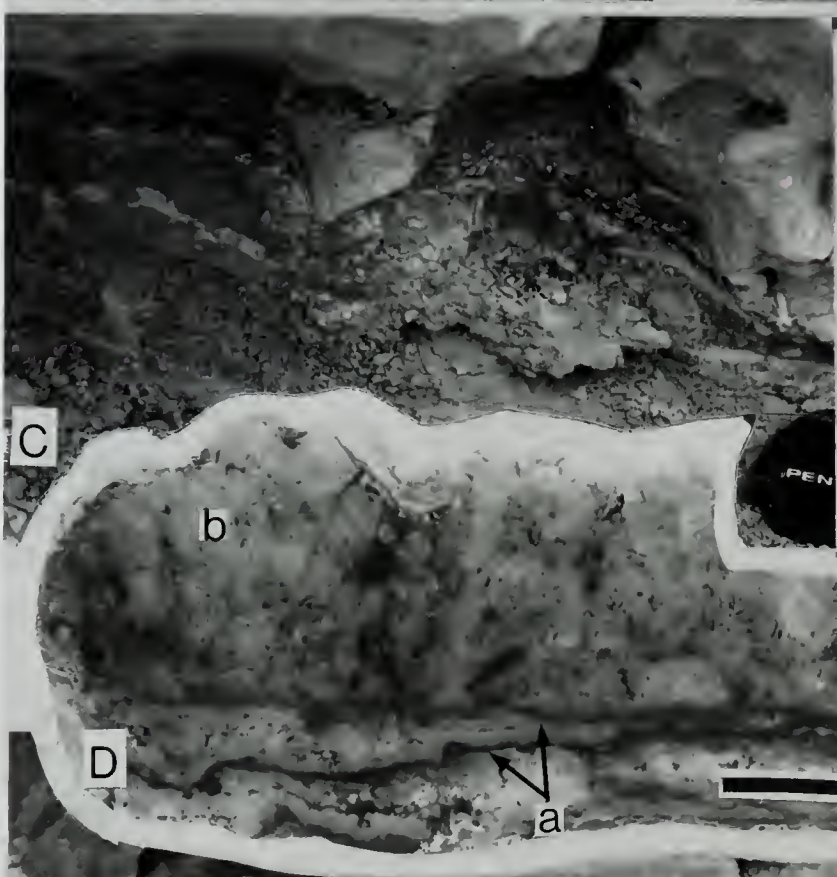
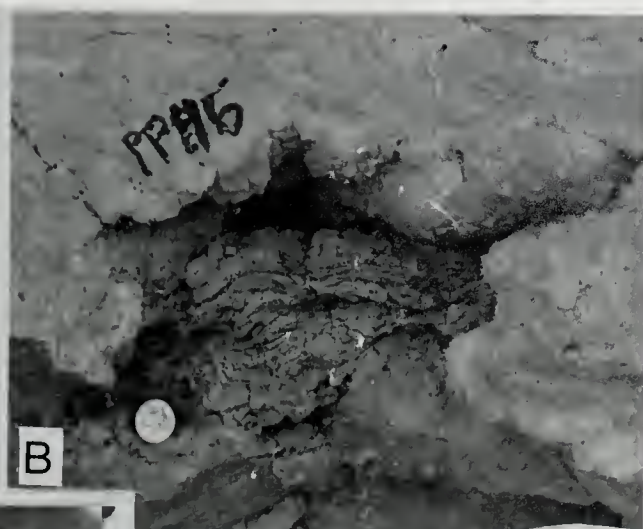
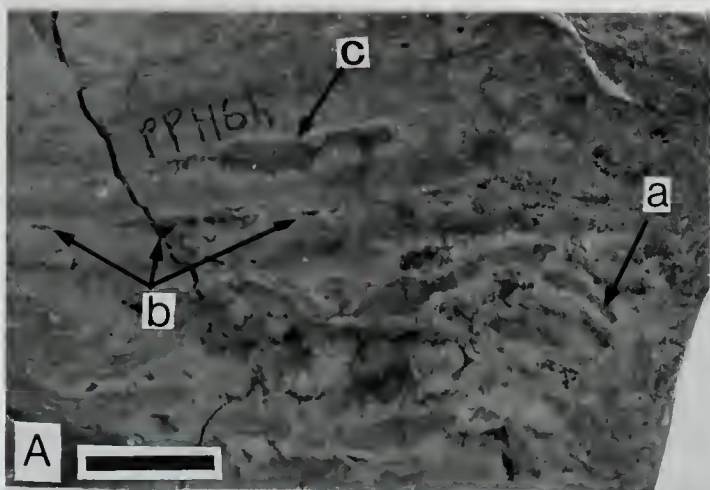
22B) Field photograph of a large vug (cave) in lithofacies 2b that is infilled with shale of the Peace Point Member (section PH). Note that the shale is laminated. Dime for scale.

22C) Field photograph of laminated shale of the Peace Point Member that has infilled a vug in lithofacies 2b. Vug formed prior to deposition of the shale. Lens cap for scale.

22D) Hand specimen of highly aggraded lithofacies 2b. Non aggraded, brown limestone occurs as relict laminations that are laterally continuous across slab (a); however aggraded sparry calcite (b) comprises most of the slab lithology. Scale bar equals 2 cm.

22E) Hand specimen of highly aggraded lithofacies 2b. Relict brown limestone patches (a) occur throughout the slab. E) underwent a higher degree of aggradation than D). Scale bar equals 2 cm.

22F) Field photograph of rosette calcite lined vug in lithofacies 2b. Scale bar equals 2 cm.



Lithofacies 2b is comprised mainly of a variably aggraded, clear to dusty, medium crystalline to extremely coarse crystalline limestone. In outcrop, subhedral to euhedral calcite crystals are elongated subvertically to vertically. Radiating scalenohedral calcite forms are rare. Crystal boundaries are generally straight and planar, however sutured contacts among crystals do occur, commonly in association with the finer aggraded calcite. Calcite forms up to 70 to 80% of the lithofacies and is generally interlaminated/interbedded with brown argillaceous calcite bands (Figs. 22D, 22E). These bands represent the relict portions of the original limestone. The calcite in the bands ranges from dense, opaque micritic rinds to 20 micron wide zoned calcite rhombohedrals. Zonation of the crystals occurs where argillaceous micrite centers are rimmed by a cleaner, coarser calcite. Overgrowths of calcite commonly surround the zoned rhombs. The aggradation of the calcite and the abundance of the zoned rhombs increase and decrease from the center of the bands, respectively.

Clear dolomite occurs as a replacement along the outer edges of the zoned calcite rhombohedrals and in amorphous, 0.5 mm diameter clusters in the calcite.

Throughout lithofacies 2b, but especially at section PH, bands of laminated green mudstone occur lining vugs (Figs. 22B, 22C). The bands thin, thicken,

pinch out around highs in bedding and are laterally continuous for a maximum length of 5 m. Lamination thickness is variable, but generally less than 2 to 3 mm.

Commonly, 1 m long, lenticular, calcareous sandstone lenses occur (section PE, east end). These lenses represent Cretaceous sandstones that were washed through, and subsequently deposited in, lithofacies 2b (Fig. 23A).

LITHOFACIES 2c (LIMESTONE/INTERMIXED DOLOSTONE)

Partial to complete replacement of lithofacies 2a and 2b produced a prominent, pale red to brown, more commonly orange, weathering, pale brownish buff freshly colored dolostone. Dolomite replacement of the limestone (lithofacies 2a and 2b) is a gradational process consisting of concordant and discordant replacement.

Concordant replacement of limestone, by dolostone, almost always results in complete replacement. Such replacement is typically associated with the massive bedded dolostones where interbedded/interlaminated dolostone and limestone occur. These rocks will be referred to as lithofacies 2c (massive dolostone variety). Generally, beds are planar with a 2 to 3 cm average bed thickness. Massive bedded (concordant) dolostone consists of two distinct dolomite types. Generally, 40% of the lithofacies comprises a 5 to 12 micron wide anhedral dolomite and 45% < 4 micron wide

FIGURE 23

23A) Field photograph of a Cretaceous calcareous sandstone lens (a) in lithofacies 2b. Scale bar equals 0.5 m.

23B) Field photograph of chalky, white, porous dolostone /limestone (a) and lithofacies 7 (b) in lithofacies 10. From a collapse structure in section PG6. Boundary between the two units is a "zig-zag" fault pattern.

23C) Field photograph of a rounded clast of lithofacies 7 in the shale matrix of lithofacies 10. From a collapse structure in section PG6. Scale bar equals 2 cm.

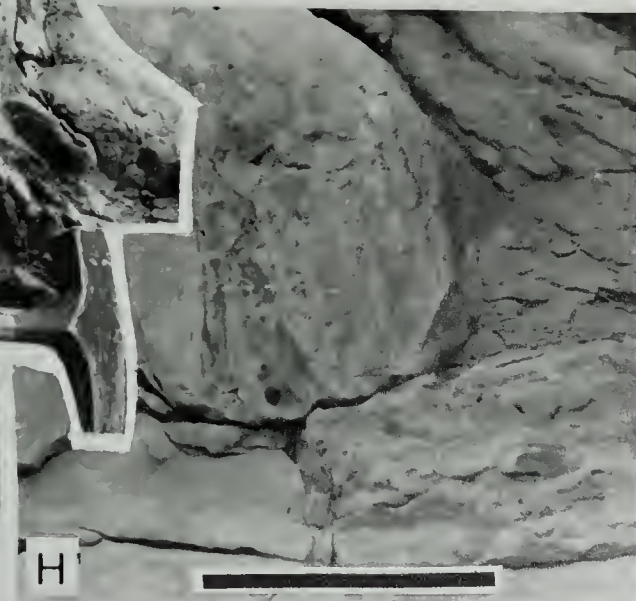
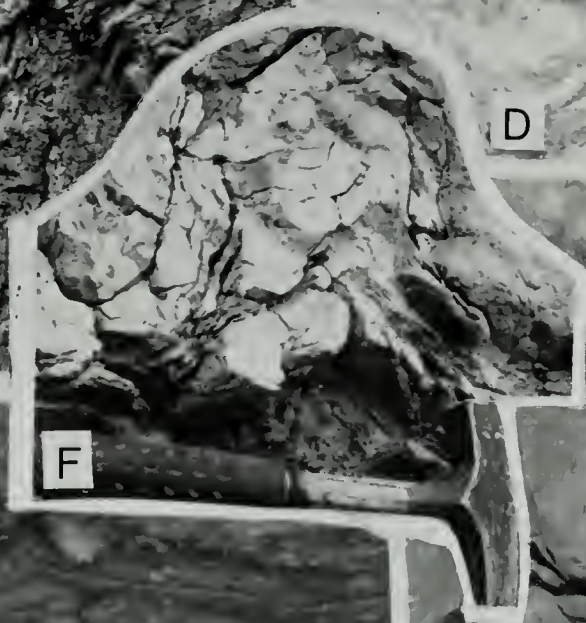
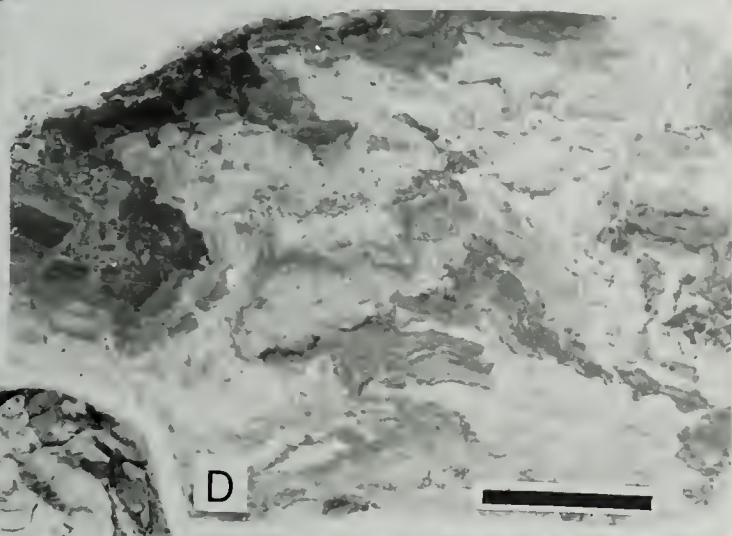
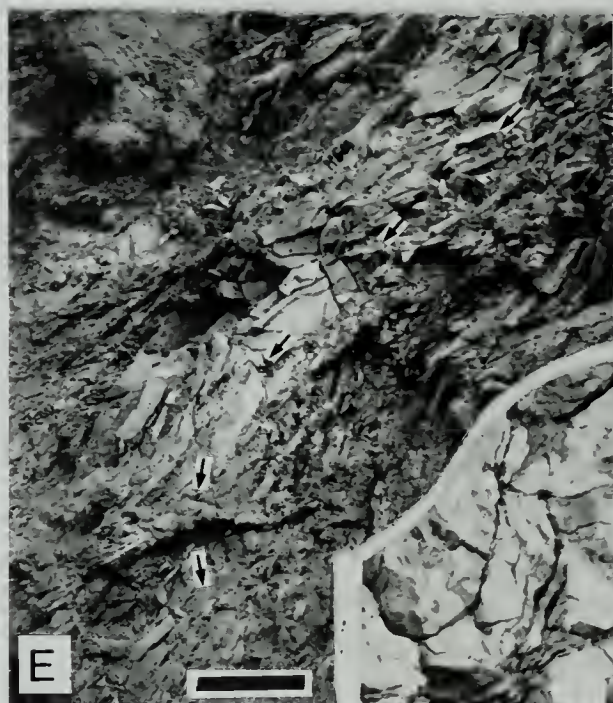
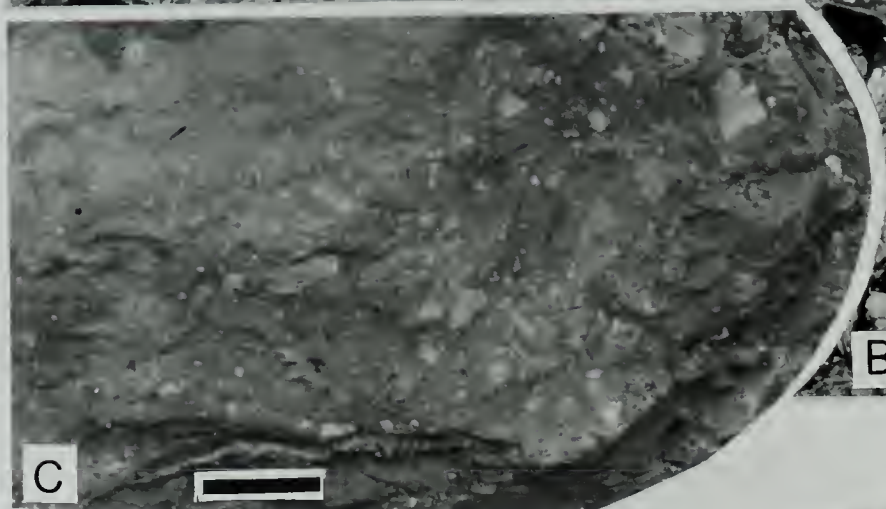
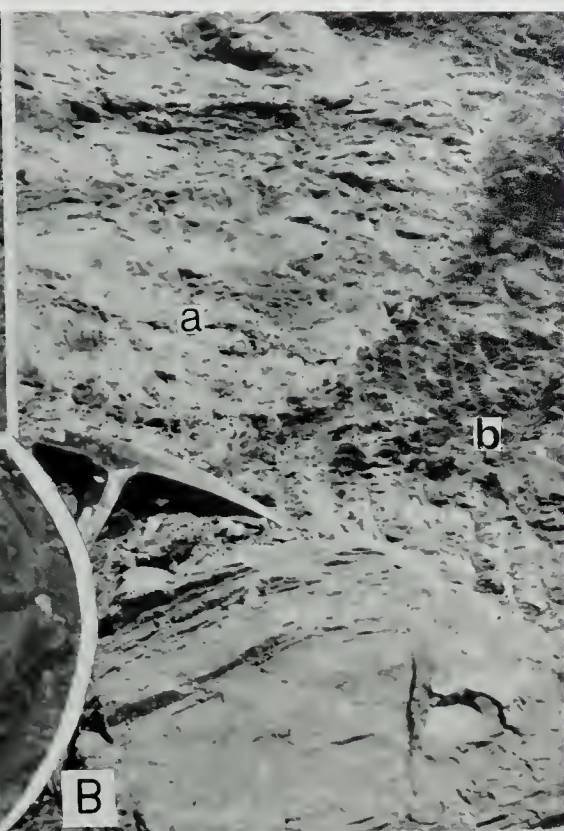
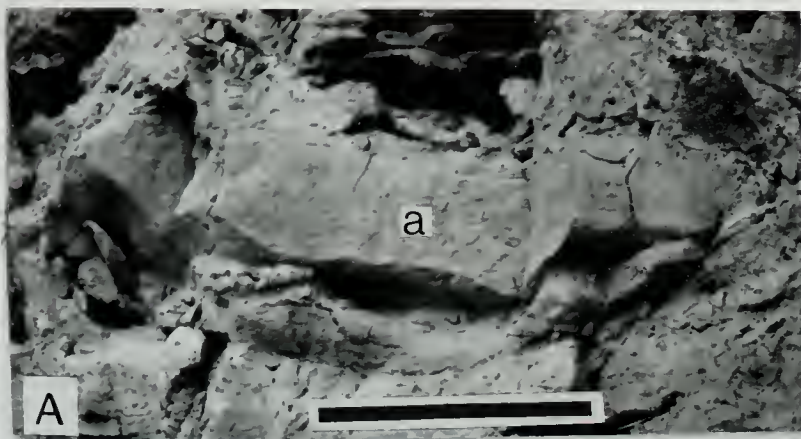
23D) Hand specimen of lithofacies 2d (section PK) showing the ropey texture and convoluted folds developed in the interlaminated/bedded dolostone (light areas) and limestone (dark areas). Sample taken from the edge of a collapse. Scale bar equals 2 cm.

23E) Field photograph of lithofacies 2d (section PG3) from western edge of collapse containing lithofacies 10 (section PG2). Arrows denote the direction of bending of lithofacies 2d. Collapse occurs to the left. Scale bar equals 10 cm.

23F) Field photograph of anticline in lithofacies 2d (section PG3). Photograph taken 1 m above Figure 26E.

23G) Hand specimen of lithofacies 2d (section PK). The lithofacies is stretched to the point of being brecciated. Scale bar equals 2 cm.

23H) Field photograph of lithofacies 2d showing the deformed texture of lithofacies 2d in section PK. Scale bar equals 10 cm.



anhedral dolomite. Together the two dolomites form a clotted texture. This lithofacies also comprises up to 20%, laterally continuous to wispy, 1 mm thick, argillaceous laminae, < 5% organics and iron dolomite, < 3% polycrystalline quartz and 1 mm thick, linear, randomly oriented, calcite veinlets.

Discordant replacement represents the initial stage of replacement in the Slave Point Formation. The replacement occurs as finger-like extensions, veinlet-type intrusions or flame-like structures of dolostone into limestone (Figs. 24A, 24C). Commonly, one lithology rims the other (Figs. 24D, 24E). With continued replacement, a bedded, concordant, dolostone may result (Figs. 24B, 24C). Rocks displaying a discordant texture will be referred to as lithofacies 2c (mixing variety).

The limestone in lithofacies 2c is dark brown, dense and opaque. It is comprised of a 12 micron wide anhedral calcite that can be aggraded to a subhedral to euhedral 76 micron wide calcite. Floating textures are produced where the 76 micron wide calcite surrounds the 12 micron wide calcite. The coarser calcite is clearer and occurs as aggraded patches and channels throughout the limestone. Generally, these patches and channels are abruptly terminated against the dolostone patches, but where the calcite enters the dolostone, it is extensively replaced. This, along with the fact that

FIGURE 24

24A) Hand specimen of lithofacies 2c (mixing variety) with limestone (dark areas)

"breaking up" in itself along stylolites (a). Dolostone (light areas) is concentrated along stylolites (b). Argillaceous laminae (clay cutans) in the dolostone follow the contour of the dolomitizing front (advancing dolomite front shown by small arrows). Scale bar equals 2 cm

24B) Hand specimen of lithofacies 2c (massive dolostone variety). Limestone has been completely replaced by dolostone. Scale bar equals 2 cm.

24C) Field photograph of lithofacies 2c (massive dolostone variety) from section PX. Dolostone (light area) has replaced limestone (dark area) with the replacement process having moved downward through the limestone, producing finger-like extensions of dolostone replacement (a). The dolostone replacement process produces a massive, concordant dolostone that assumes the same bedding character, bedding thickness and texture of the previous limestone. Lens cap for scale

24D) Field photograph of lithofacies 2c (mixing variety). Photograph was taken below the base of the cliff containing section PL. A mixing process where small dolostone clasts (a) occur in a limestone that is breaking up internally (as 27A). The chemical mixing of the limestone and the dolostone formed lithofacies 7 in the center of the photograph (c).

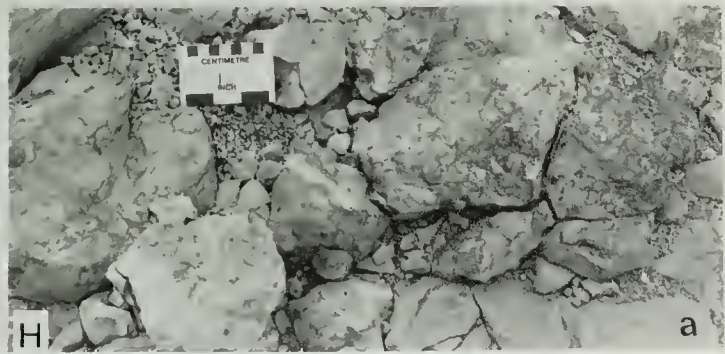
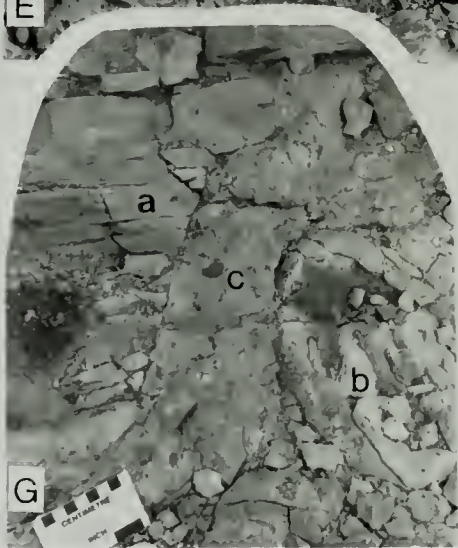
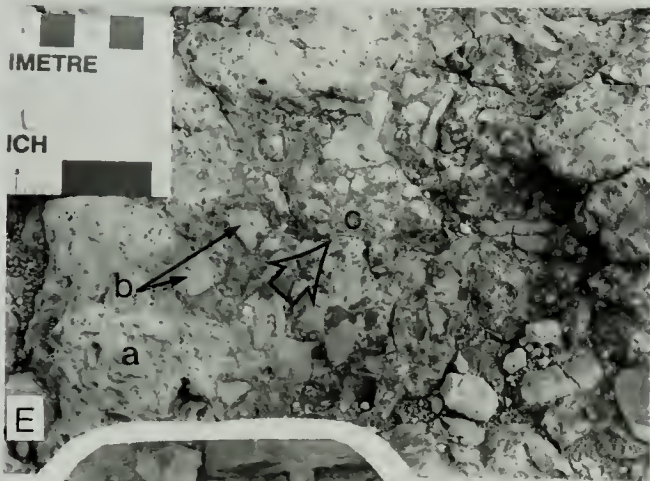
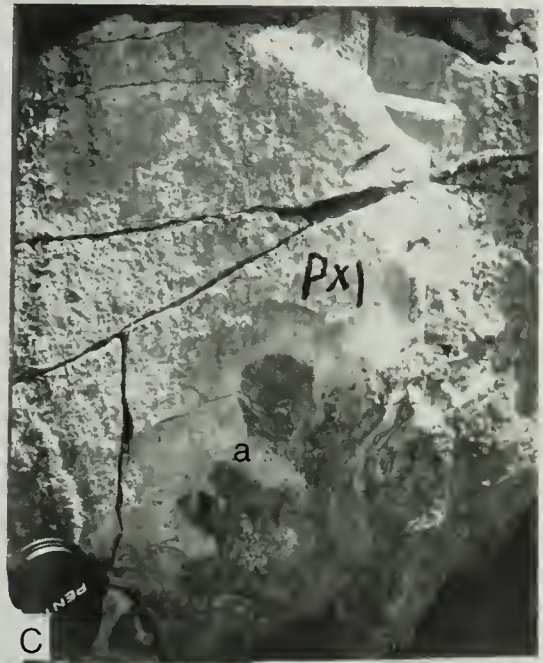
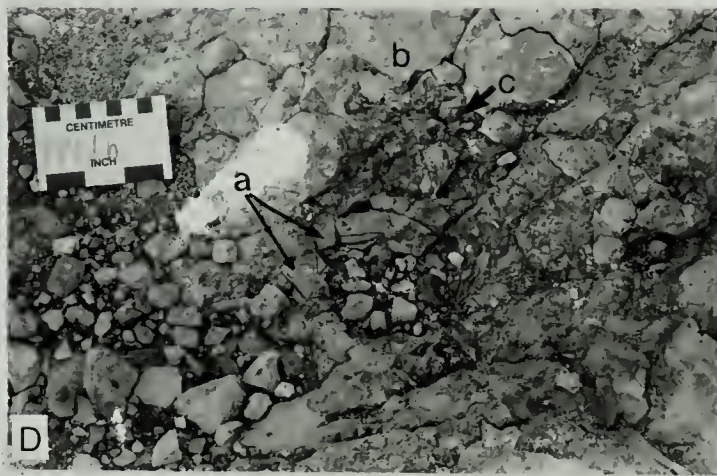
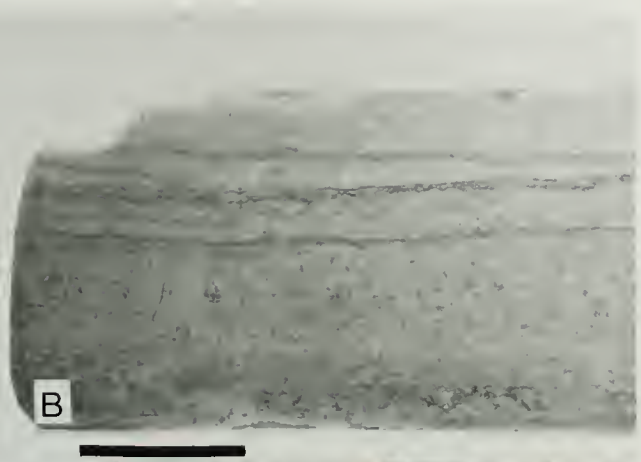
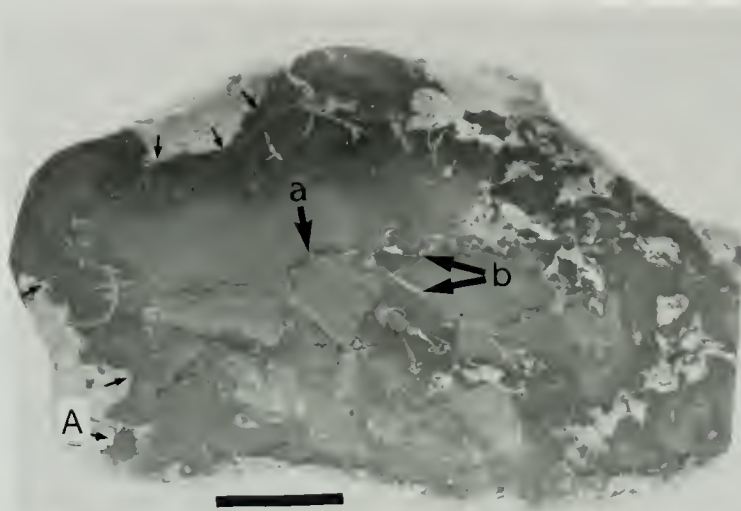
24E) Field photograph of lithofacies 2c (mixing variety). Limestone which breaks up internally (a) forms a surrounding rock type. The chemical mixing of dolostone and limestone of lithofacies 2c(b) has formed lithofacies 7(c). With an increase in chemical mixing the dolostone became more brecciated. The arrow points in the direction of a gradational change from limestone, through lithofacies 2c to lithofacies 7.

24F) Field photograph of a block of lithofacies 2c (massive dolostone variety, a) that underwent chemical dissolution and was later incorporated into a breccia (b) that is comprised solely of homogenic limestone clasts. The breccia is similar to lithofacies 2c (mixing variety), but in this case, appears to be a younger event than (a). Dissolution of (a) and (b) is obvious along a highly irregular contact (d). This mixing process forms a breccia (b) that is extremely similar to the matrix of lithofacies 8 seen elsewhere in the study area.

24G) Field photograph showing lithofacies 2c (massive dolostone variety, a) and

lithofacies 2c (mixing variety, b) cut by a younger breccia pipe (c). Note the sharp contacts on either side of the pipe. The breccia pipe contains both varieties of lithofacies 2c as well as shale of the Peace Point Member. This pipe is identical to the matrix of lithofacies 8 seen elsewhere and represents dissolution that was associated with the collapse forming lithofacies 8. This photograph was taken from section PQ, where lithofacies 8 underlies the photograph. This photograph shows a less brecciated, less chaotic texture of lithofacies 8.

24H) Field photograph of lithofacies 2c (mixing variety). Note how the dolostone becomes more homogeneous towards the center of the block (a) with mixing more obvious towards the outer edges of the block



both the 12 micron and 76 micron wide calcite are corroded by the dolomite, suggests that dolomitization was the younger event.

Dolomite patches in the limestone are concentrated and generally coincide with the stylolites (Fig. 24A). In some samples, the limestone of lithofacies 2c is divided by stylolites, thus producing a brecciated texture (Fig. 24A). The stylolites are lined with bright orangey brown and reddish brown clay cutans and 12 micron wide dolomite. The initial stage of dolomitization of lithofacies 2 is somehow related to the stylolites. Dolomitization occurred as an advancing front which moved in both directions away from the stylolite, eventually replacing the limestone (Fig. 24A). The bright colored argillaceous laminae (cutans) form parallel to the stylolitic contact between the dolomite and the limestone. These clay cutans are either insoluble residues inherent in the limestone, that were not consumed by the dolomitization process, or they represent a product of the dolomitization process. The former theory is more likely because as the dolomitization front proceeded the insoluble clays were left behind to form laminae in the dolostone. This would also account for the conformable nature of the clay cutans to the dolomitizing front. Limestone that occurs adjacent to the dolomitization front is generally bleached, thus suggesting that replacement was already

occurring in the limestone.

The 12 micron wide dolomite forms a clotted texture with dolomicrite. As replacement of the limestone progressed, the dolomicrite content increased substantially, thus suggesting that the dolomicrite is probably a degradation of the 12 micron wide dolomite. The dolomicrite is concentrated around the more argillaceous laminae, which may also have aided in the dolomicritization process.

LITHOFACIES 2d (CONVOLUTED/ROPEY LIMESTONE/DOLOSTONE)

Lithofacies 2d typically reflects the partial replacement of limestone by dolostone and has the added feature of being physically stretched, convoluted and in some cases brecciated (Figs. 23E, 23F, 23G, 23H). This lithofacies forms a semiprominent to prominent, white to medium brown (commonly buff) dolostone and interlaminated, interbedded light grey to black limestone. The dolostone contains < 1 mm thick light brown to black argillaceous laminations/bands that are always parallel to the relief of the interlaminated limestone. The dolostone completely cuts across limestone clasts in the brecciated portions of lithofacies 2d, thus suggesting that it is a younger event.

The limestone of lithofacies 2d is considerably darker than in other lithofacies, with the center of the limestone bands always darker than the outer portions.

Since the dolostone may extend into the limestone a nodular, but bedded unit may result. The limestone of lithofacies 2d is denser and more finely crystalline than the limestone in lithofacies 2a to 2c. Rarely, dark, indistinct, thin, featureless dense micritic rinds coat some of the limestone clasts, which can in places be quite soft.

Lithofacies 2d is internally gradational. The stretched nodular texture of this lithofacies may pass laterally into a highly contorted, convoluted lithofacies where asymmetrical folds and microfaults occur (Fig. 23D). With increased horizontal strain the lithofacies changes to a breccia with a rock type ranging from a particulate crackle floatbreccia to a particulate rubble floatbreccia (Fig. 23G). Generally, the clasts are elongated along one axis, but with increased deformation they become more spherical. The clasts are comprised of black limestone and dolostone, whereas the matrix is a medium brown color formed from the mixing of the two clast types. From cut and slabbed portions of rock from this lithofacies minor pockets of breccia occur overlain and underlain by laminated rock, thus suggesting collapse on a microscale. Increased disruption of lithofacies 2d produces a darker color, a lower argillaceous content, and a higher dolomitic content.

DISTRIBUTION AND THICKNESS

Lithofacies 2b and 2c are the most laterally extensive, whereas lithofacies 2a and 2d are laterally restricted. Areas comprised solely of lithofacies 2a are rare, but lithofacies 2a does occur as an isolated remnant of unaltered material in lithofacies 2b (section PT) and below the cliff containing section PB. Lithofacies 2d is generally restricted to those areas adjacent to the larger breccia bodies. Patches and minor intermixing of all four lithofacies does occur throughout the study area, however, only the more extensive and more homogeneous areas have been shown (Fig. 23).

The discordant type of replacement is best observed directly overlying small gypsum anticlines (lithofacies 1) where lithofacies 2c generally comprises a brown limestone and light brown to buff dolostone (Fig. 21C). There, the dolomite replacement is patchy, with only partial replacement of the Slave Point limestone having occurred. Dolomitization is minor on the very edges of the small gypsum anticlines, whereas, the larger, more open folds (section PE) are generally unconformably overlain by a distinctly bedded, well developed, concordant dolostone. Lithofacies 2b or 2c is generally bedded parallel to the unconformable surface that occurs at the base of the Slave Point Formation. Lithofacies 2b generally overlies lithofacies 2c. Where lithofacies 2b

is extremely vuggy (sections PE and PH), lithofacies 2c is developed as a concordant, well bedded dolostone beneath lithofacies 2b (Fig. 21A). However, where lithofacies 2b is not as vuggy, lithofacies 2c only occurs as a discordant, patchy replacement of the limestone beneath lithofacies 2b (Fig. 21C). In other areas the patchy aspect of the dolostone replacement is not as evident and lithofacies 2b lies directly on top of lithofacies 1.

The upper contact of lithofacies 2b is either gradational with lithofacies 2d, 7 or 8 or displays a sharp, slightly wavy to planar contact with lithofacies 3.

Overall, lithofacies 2 is gradational laterally into lithofacies 7 and 8. The gradation is recorded as an increase in brecciation and an increase of mixing between the limestone and dolostone lithologies (Fig. 25).

Lithofacies 2b and 2c reach maximum thicknesses in section PE and section PH. In section PE, 5.5 m of lithofacies 2c (massive dolostone variety) is overlain by 5.25 m of lithofacies 2b. There, the bedded strata of the Slave Point Formation attain a maximum thickness of 11.75 m in the study area. In section PH 8.35 m of bedded Slave Point occur with the lower 7.6 m representative of lithofacies 2. Of that 7.6 m, the lower 5.2 m comprises lithofacies 2c and the upper 2.4 m

FIGURE 25

25A) Lithofacies 2a represents the most original, unaltered limestone in the Slave Point Formation (A). This lithofacies became either highly aggraded, producing lithofacies 2b (B) or was replaced with dolostone, producing lithofacies 2c (B'). The A-B-C-D series represents a limestone to calcrete series that was eventually incorporated into (E) whereas, the B'-C'-D' series represents chemical mixing and dissolution more typical of karst. The B'-C'-D' series is basically a change from the mixing variety to the massive dolostone variety of lithofacies 2c with subsequent brecciation having occurred. Both series are incorporated into lithofacies 8. All scale bars equal 2 cm.

25B) Lithofacies 2b represents a highly aggraded limestone (white areas) that is pisolitic in places. With continued subaerial exposure (C) was produced.

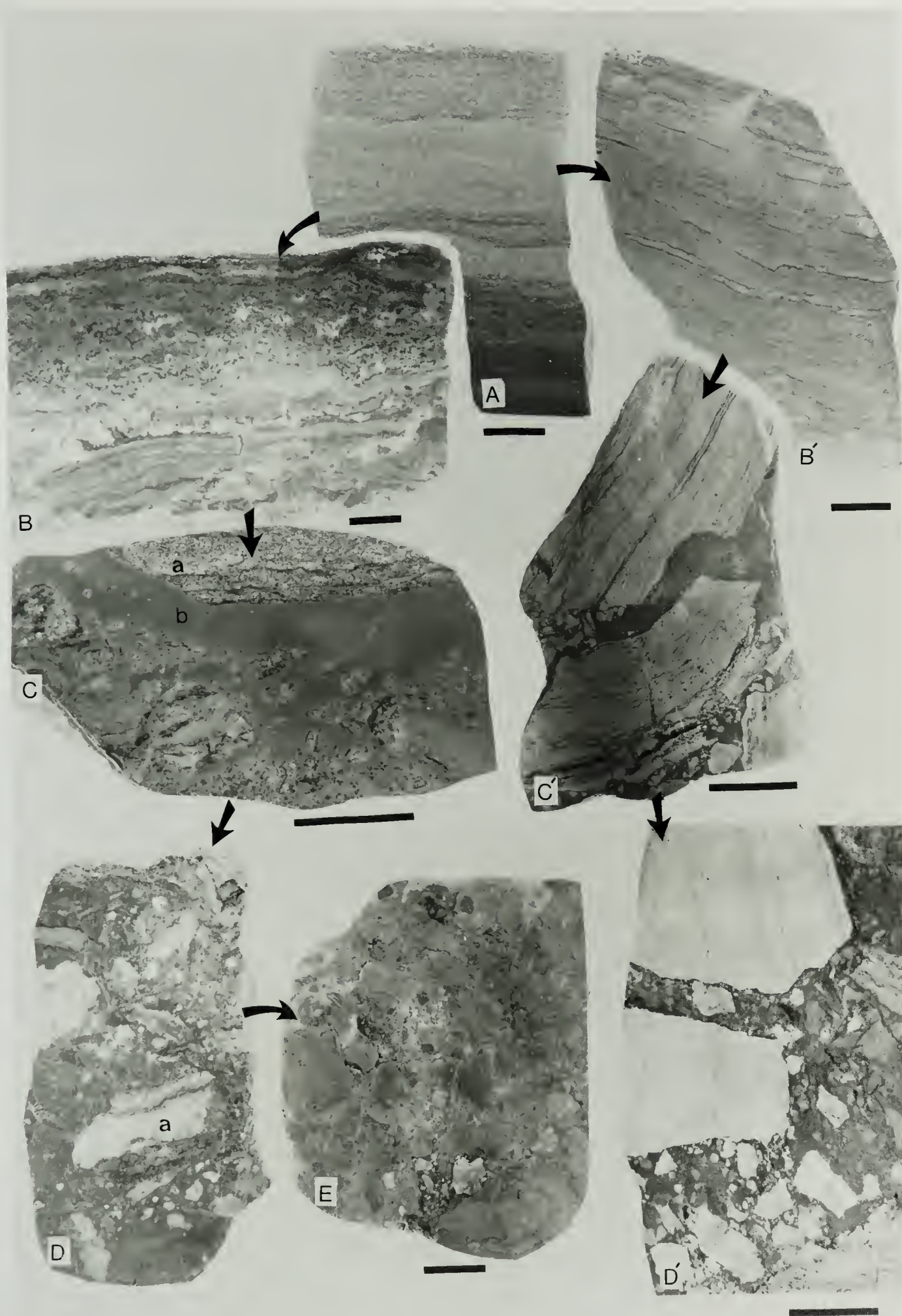
25C) Calcrete portion of lithofacies 8. The slab contains features typical of calcrete that represent further diagenesis of lithofacies 2b. The matrix becomes sugary (a) with replacement by dense, opaque micrite (b) having occurred.

25D) Calcrete breccia? (from lithofacies 8). (C) became more chaotic, brecciated and developed a higher degree of alteration. The limestone clasts (a) are more diagenetically altered than in (C, a).

25E) Upon collapse the calcrete portion of (C) was mixed with shale and numerous other lithofacies to form lithofacies 8.

25B') Lithofacies 2c represents the massive dolostone variety (shown) produced by a gradational dolomitization process (mixing variety, not shown).

25C' and D') Lithofacies 2c was brecciated and incorporated into lithofacies 8 (C') and in areas where brecciation was more extensive lithofacies 2c was mixed in with shale and other lithofacies forming lithofacies 8 (D').



comprises lithofacies 2b. These thicknesses vary laterally because of the manner in which the dolostone has replaced the limestone.

LITHOFACIES 3 (GYPSUM MOULD BRECCIA)

Lithofacies 3 is a prominent, pale reddish pink to pale greenish buff weathering, light buff to light brown fresh surface colored dolostone. Bedding thicknesses range from 1 to 8 cm, with an average of 4 cm. The dolostone is interlaminated with light buff and light brown laterally continuous laminations that range from less than 1 mm to 3 mm thick. Ellipsoid-shaped, rounded to well rounded moulds, that are filled with a clear sparry calcite cement, form 15% of the lithofacies (Fig. 26A). The moulds are up to 6 mm long, with an average length of 2 mm. They generally form parallel to and along the laminations. The moulds may represent dissolved gypsum moulds that were later filled with a clear, sparry, calcite cement. Clear calcite cement also lines 0.5 mm wide fractures that can form up to 20% of the lithofacies.

Lithofacies 3 contains 1 to 2 cm wide localized areas of collapse breccia. This, along with the presence of relict gypsum textures and the gradational aspect of the breccias in this lithofacies, suggests that the breccias are of a collapse origin. Lithofacies 3 changes from a laterally laminated, non brecciated dolostone, to a cemented crackle packbreccia, to a particulate mosaic packbreccia.

FIGURE 26

26A) Hand specimen of lithofacies 3 showing laminated dolostone with isolated areas of collapse (a). Oval to round moulds are infilled with white cloudy dolomite (b). Scale bar equals 2 cm.

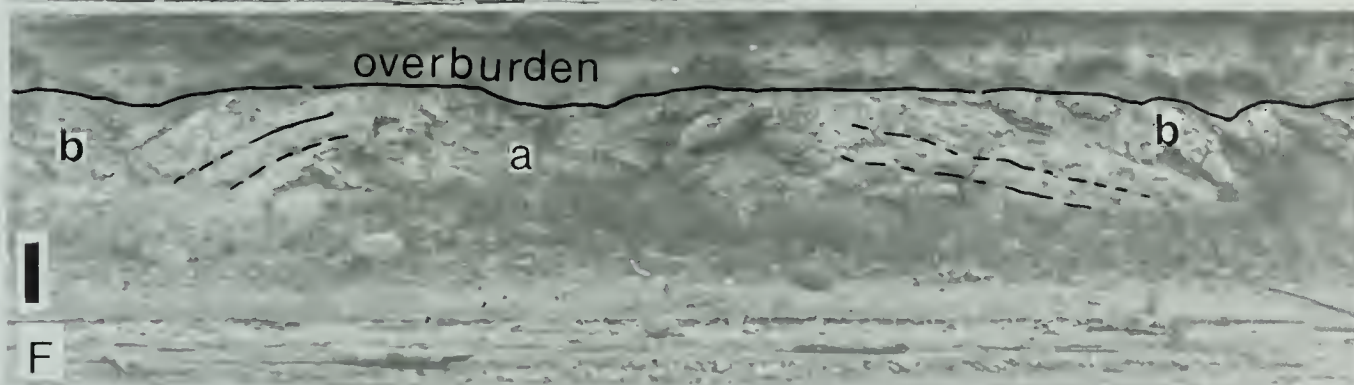
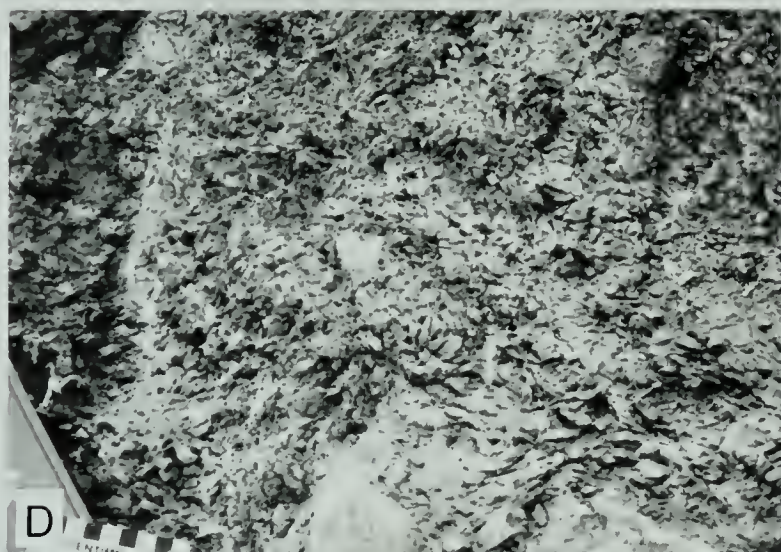
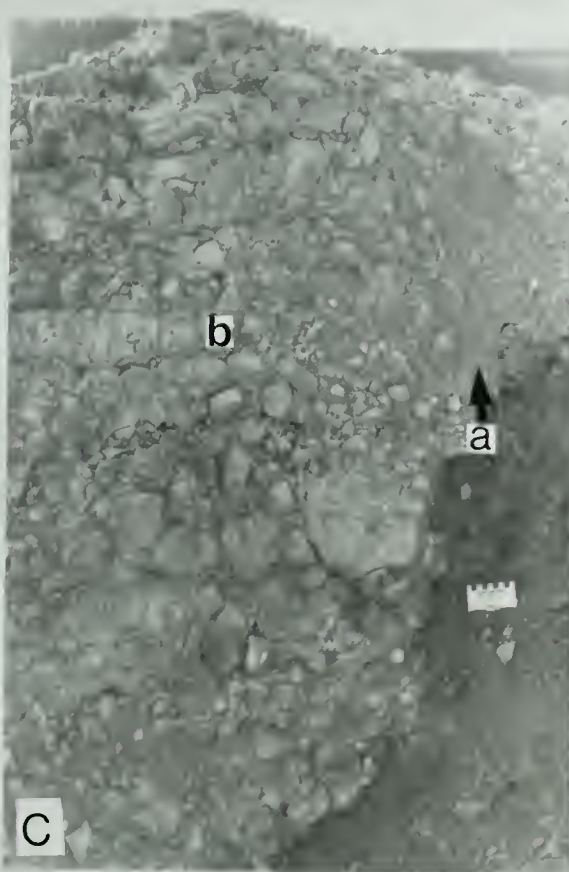
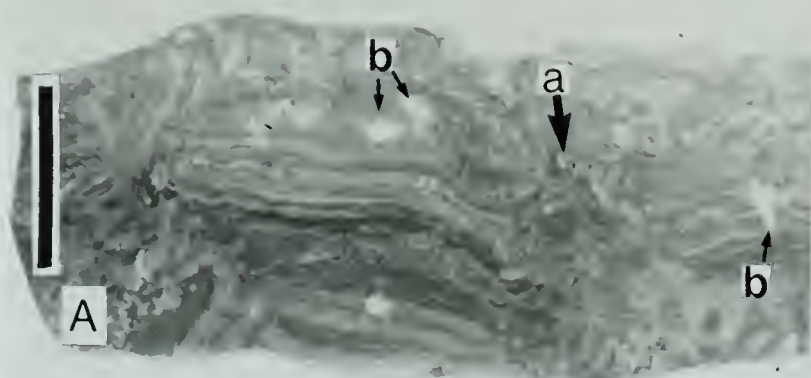
26B) Field photograph of lithofacies 4 (section PM) showing the blocky texture and restricted lateral extent of the breccia. Scale bar equals 1.5 m.

26C) Field photograph of lithofacies 4 (section PM) showing the restricted width of the breccia in the third dimension. Note the gradational contact (a), relict bedding (b), the overall homogeneity of clast size and the angularity of the clasts.

26D) Field photograph of lithofacies 5 showing the chaotic and friable texture of the shale.

26E) Field photograph of lithofacies 5 (section PC) in a pocket in lithofacies 8. The dashed line defines the extent of the pocket.

26F) Field photograph of lithofacies 6 showing a relict domal shape. Brecciation has occurred at the crest (a) and in the synclines adjacent to the dome (b). Scale bar equals 1.5 m.



DISTRIBUTION AND THICKNESS

Lithofacies 3 occurs only in section PH where 0.75 m of it overlies lithofacies 2b with an abrupt, planar to slightly wavy contact. The upper contact is not seen in the study area. Laterally the lithofacies bends down into a breccia zone that occurs on the west side of section PH. Lithofacies 3 is laterally and vertically restricted and when its spatial distribution is taken into consideration it is rather an insignificant unit; but it is lithologically unique and is therefore regarded as a lithofacies.

LITHOFACIES 4 (BLOCKY BRECCIA)

Lithofacies 4 forms an extremely prominent brecciated unit that weathers white, creamy white, brown, and medium grey with minor orangey rust staining (Fig. 26B). The breccia is almost lithologically homogenous with 70 to 90% of the clasts comprised of a laminated, non dolomitic to dolomitic, pelletal, aphanocrystalline limestone. Laminae of micrite and pelmicrites form interlamination in the limestone clasts. The micrite laminae are argillaceous, medium brown and laterally continuous across the clasts, whereas, the pelletal micrite laminae are light greyish brown and contain 70 to 90% pellets. Blocky calcite cement forms around the pellets and can comprise 15% of the clasts. There is minor aggradation of the micrite and aggraded channels of calcite cut across laminae in this clast type. These clasts are generally replaced with an

aphanocrystalline dolomite, however, crystals can attain a few tens of microns in width and up to 15 to 20% of some clasts. Clasts contain, 5 to 15% bioclastic debris, < 5% crinoid ossicles, thin shelled brachiopods (non-articulated), ostracods and trace amounts of quartz.

A second clast type in lithofacies 4 is a massive brown limestone clast that is similar to the previous clast type, however, these clasts contain no internal lamination, and only contain 10 to 20% pellets. The clast lithologies mentioned above do occur in other breccias, but not to the extent that they are found in this lithofacies. These clasts were not found in the bedded Slave Point strata, but possibly represent some original unaltered rock of the Slave Point Formation for they are similar in description to the laminated, light to medium brown, micritic to fine grained, fissile limestone described by Norris (Norris and Uyeno, 1983).

A third clast type is a breccia that comprises lithologies of the two previously described clasts. Pelmicrite and micrite particles in the breccia clasts are floating in a matrix comprised of 15 micron wide anhedral calcite. The breccia clasts also contain euhedral rhombs of calcite, amorphous blebs of chert, isolated silica/quartz crystals (10 to 15%) and iron dolomite (up to 5%).

Fourth and fifth clast types are the mixing variety and massive dolostone variety of lithofacies 2c. These clasts form up to 10% and < 5% of the clasts, respectively.

Clasts in lithofacies 4 are < 1.5 m in length with a bimodal size distribution consisting of 4 cm and 20 to 24 cm modes. Lithofacies 4 displays a high clast to matrix ratio, generally about 80:20. This clast supported texture and angular shape of the clasts cause the breccia to have a "blocky" physical appearance (Figs. 26B, 26C).

The brown argillaceous matrix of the breccia is always the most recessive part of the breccia. The breccia is best named a particulate rubble packbreccia.

There are a number of characteristics of lithofacies 4 that delineate it as an unique lithofacies, namely:

1. It has a restricted lateral extent.
2. It is a packbreccia with a high clast to matrix ratio.
3. It has a chaotic texture with very angular clasts.
4. It is predominantly a pelletal micrite lithology.
5. There are no shale clasts in it.
6. There are rare dolostone clasts in it.
7. There is only the rare occurrence of clasts of lithofacies 2c.
8. There are clearly 2 phases of brecciation. Clasts in the breccia are comprised of breccia.
9. The clasts are clearly distinct from the matrix.

DISTRIBUTION AND THICKNESS

Lithofacies 4 is found in sections PU, PT, PN and PM. In section PT, lithofacies 4 attains a maximum thickness and lateral extent of 10 and 9 m, respectively. In sections PM and PN the breccia body is

visible in three dimensions, where the width (in towards the cliff) is approximately 1.5 m. The north facing wall of the breccias are defined by a gradational contact (over 0.5 m) where the more prominent blocky breccia of lithofacies 4 changes to the recessive, argillaceous, finer grained area of the breccia (lithofacies 9) and cliff talus behind (Fig. 26C). Lithofacies 4 generally occurs as vertically elongated breccia pipes that are ellipsoidal to rectangular in cross section (perpendicular to elongation).

The lower and upper contact of lithofacies 4 are not exposed in the study area. Laterally, the breccia appears to cut through both lithofacies 8 and 9.

LITHOFACIES 5 (SHALY BRECCIA)

Lithofacies 5 is a recessive, shaly breccia with a matrix consisting of bluish bright green shale and minor amounts of brown shale. The fact that the lithofacies is highly recessive makes it difficult to discern it from talus or scree developed along the cliffs. However, where discernible the shale is highly contorted, friable, soft and calcareous (Fig. 26D). White sparry calcite precipitated in the matrix can reach up to 15%. At section P0 a brown to greenish brown shale (0.25 m thick bands) is interbedded with 4 cm thick buff dolostone bands. These interbanded lithologies directly overlies the gypsum dome of section PP. Poorly preserved fossils are generally concentrated in orange to brick red, highly oxidized, pockets in the shale.

Bluish, light grey, Atrypa rich, bioclastic micrite clasts, that contain traces of pyrite and copper, are common. Based on the color and faunal content of both the shale and the bioclastic micrite clasts, the majority of lithofacies 5 is representative of the Peace Point Member.

The clast to matrix ratio of lithofacies 5 varies between 65:35 and 85:15. Clast lithologies include: massive brown limestone, laminated brown limestone, pelloidal aphanocrystalline limestone and clasts of lithofacies 2c and 2d. Clasts are angular and do not exceed 0.5 m in length. This breccia is a particulate rubble floatbreccia.

DISTRIBUTION AND THICKNESS

Lithofacies 5 reaches a maximum thickness of 7 m in section PY. The length of the lithofacies in outcrop is hard to determine due to the problem of separating it from surficial cover on the cliffs. Sections PA and PY are the only sections where lithofacies 5 reaches any substantial lateral extent, with a length of 50 m attained at section PA. The lithofacies also outcrops in sections PC, PY and P0.

The lower contact is irregular and abrupt. The underlying lithofacies in all cases, except one, is lithofacies 6. The exception is at section P0 where the underlying unit is lithofacies 1. There, the shale is bedded, whereas elsewhere it has a chaotic texture.

The upper contact with lithofacies 8 is an irregular but abrupt contact. In section PA, lithofacies

8 extends out over the shaly breccia in an irregular fashion. In section P0, the upper contact is with lithofacies 9. However, the nature of that contact is not apparent because of surficial cover.

Lithofacies 5 laterally changes to lithofacies 6, 8 or 9. Lithofacies 5 does not have any characteristic shape and is generally located in pockets in lithofacies 6, 8 and 9 (Fig. 26E). Maximum height and thickness of the pockets are 3.5 and 3 m, respectively.

LITHOFACIES 6 (GYPSUM BRECCIA)

Lithofacies 6 forms a semirecessive to semiprominent, white to buff weathering breccia, comprised of gypsum and dolostone clasts (Fig. 27A). The gypsum is the same as that in lithofacies 1. The orange weathering dolostone, that is characteristic of lithofacies 2c, is commonly concentrated towards the top of lithofacies 6. Dolomicrites of lithofacies 1 also occur and are distinguishable from lithofacies 2 dolostone by, a finer crystal size, less argillaceous content, thinner bedding and the presence of small oval blebs (less than 1 cm diameter) that are filled with gypsum. Small pockets of lithofacies 5, that are generally smaller than 1 m wide, typically occur.

Blocks of dolostone and gypsum generally attain maximum lengths and widths of 3 and 1.5 m, respectively; however, blocks of lithofacies 1 can attain lengths and widths of a few meters. The larger blocks (over 0.5 m) in lithofacies 6 are very angular and rectangular in shape, whereas the

smaller clasts are more rounded (Fig. 27C). Laminae in the smaller dolostone clasts are aligned parallel to the outer edges of the larger gypsum blocks, thus suggesting that the dolostone blocks had flowed around the larger gypsum blocks (Fig. 27A). The block to matrix ratio of the breccia is generally 70:30.

The matrix of lithofacies 6 weathers a greenish bluish grey with moderate amounts of orange to rust colored staining. The matrix is formed of a chaotic mixture of small particles of dolostone and gypsum in a shaly, well indurated, groundmass. Anhydritic replacement of the matrix occurs more readily than it does in the gypsum blocks. Laterally discontinuous, 3 cm wide satin spar gypsum veinlets, up to 3.5 cm thick, cut through the gypsum and dolostone clasts, as well as through the matrix (Figs. 27B, 27D). This suggests that the gypsum had been remobilized after the brecciating event. The breccia is generally a particulate rubble packbreccia.

DISTRIBUTION AND THICKNESS

Lithofacies 6 attains a maximum thickness of 14 m in section PT and a maximum length of 86 m in section P0. The lithofacies is exposed in sections PA, PB, PC, PT, PW and PY. The lower contact is not exposed in the study area. The upper contact is transitional with the overlying lithofacies 8 and is defined by the uppermost occurrence of gypsum. However, near the west end of section PA, lithofacies 5 pinches in between lithofacies

FIGURE 27

27A) Field photograph of lithofacies 6 (section PA) showing large gypsum blocks (a) in a matrix comprised of dolostone, gypsum and shale. The collapse is not as extensive as some areas because bedding on the blocks is generally the same attitude. There is some flowage around the large blocks because the smaller dolostone clasts form around the blocks with the long axis parallel to the outer edge of the gypsum. A gradational contact with lithofacies 8 is denoted by the arrows. Scale bar equals 1 m

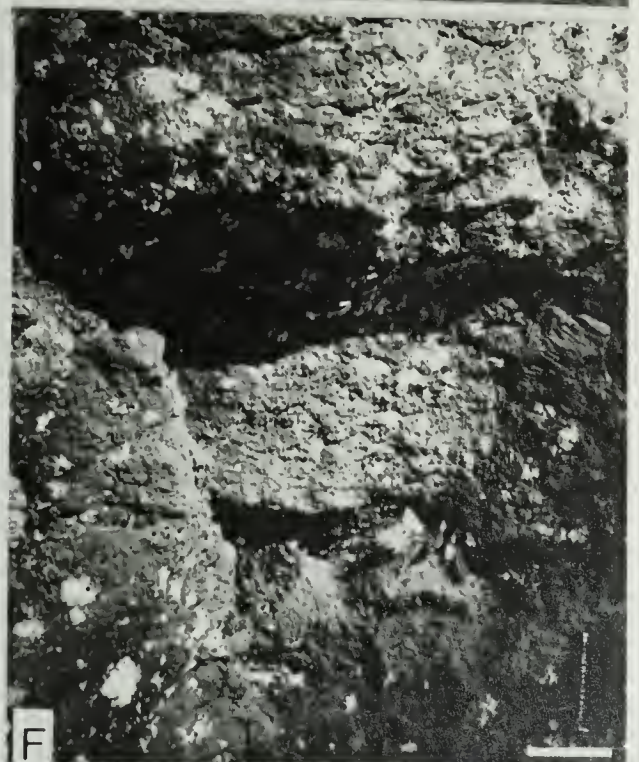
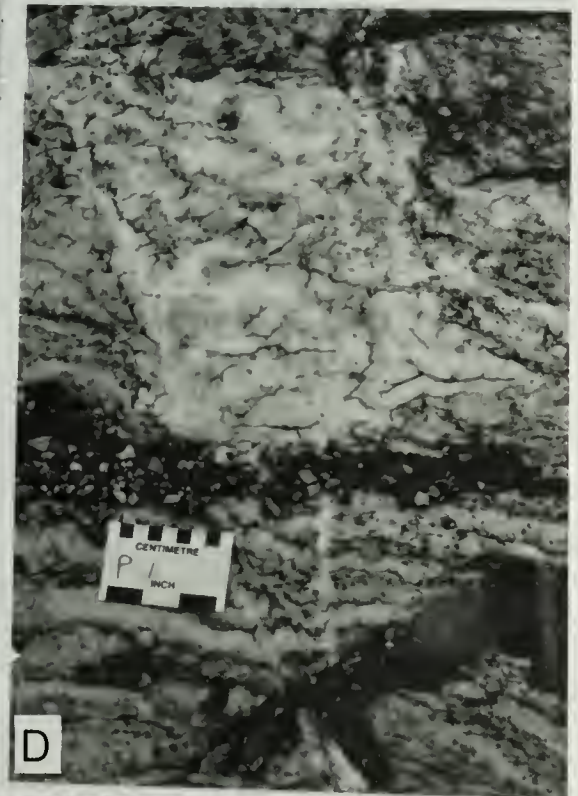
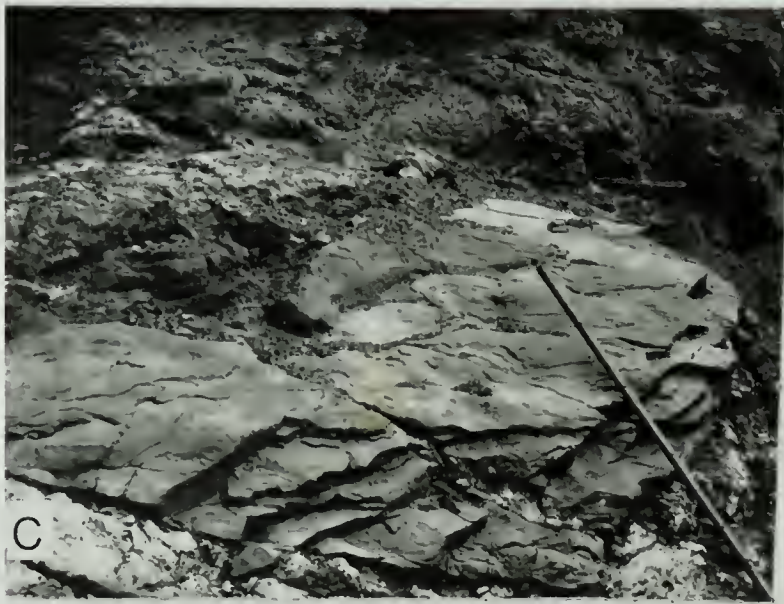
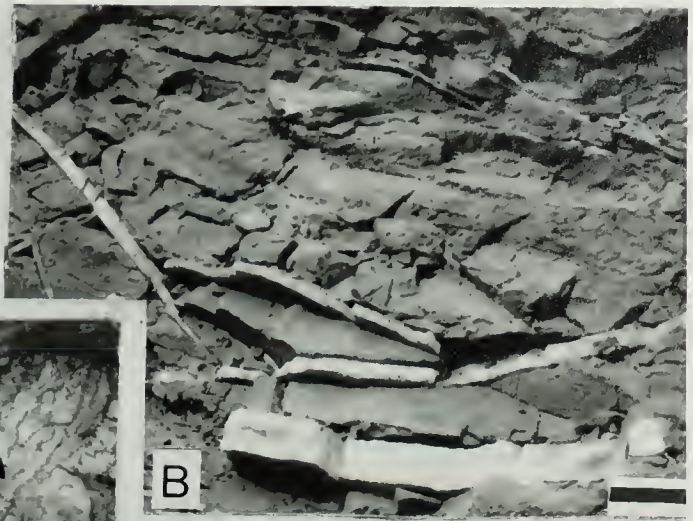
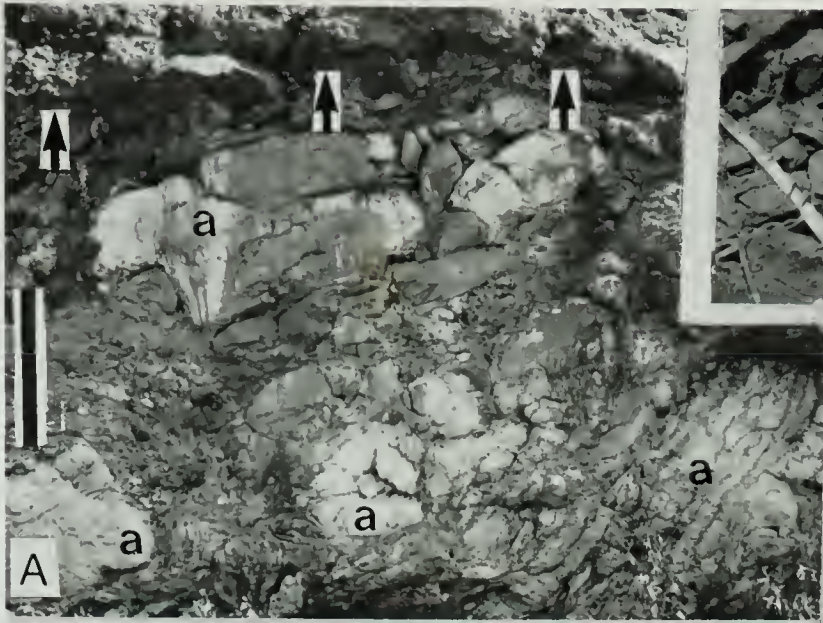
27B) Field photograph of gypsum veins cutting across dolostone blocks (section D). Veins cut across both blocks and matrix of lithofacies 6 thereby suggesting the gypsum has been remobilized since the brecciating event forming lithofacies 6. Scale bar equals 2 cm.

27C) Field photograph of large dolostone block in lithofacies 6 (section PA). Gypsum surrounds the block. Pogo stick for scale.

27D) Field photograph of gypsum veins cutting across lithofacies 1 (section PB, below the base of the cliff) thereby suggesting that the gypsum was remobilized after the deposition of lithofacies 1.

27E) Field photograph of lithofacies 7 (section PK). Soft, white, chalky, lumps of dolostone (glaebules, a) in a dark brown argillaceous matrix (b).

27F) Field photograph of Holocene terra rossa, taken on road cut approximately 4 km, west of Waitomo, North Island, New Zealand. White glaebules comprise soft, white and chalky claystone, whereas the matrix is an argillaceous limy residue. The unit is developed directly from an underlying limestone. 30E and F are identical in appearance to this Holocene example with the exception of a slightly fainter color in 30E. Bar scale equals 5 cm.



6 and 8. Lithofacies 6 extends down below the base of the outcrop and laterally terminates against lithofacies 8 (sections PE, PC, PT). This gives lithofacies 6 a characteristic domal, convex upward shape. In sections PW and PY, lithofacies 6 laterally changes into lithofacies 1. Lithofacies 6 is generally found in the lows developed between domes of lithofacies 1. In places (section PA), bedding in the larger blocks of lithofacies 6 are all dipping relatively the same direction and at the same angle, thus defining a relict homocline in the breccia. However, relict dome structures are more commonly traced through lithofacies 6 (Fig. 26F).

Lithofacies 6 displays a variation as to the degree of brecciation, with some exposures of lithofacies 6 having a more chaotic texture than others. Where bedding from one breccia block to the next can be traced there appears to have been little vertical displacement (section PA). Also, in those less chaotic exposures, dolostone blocks are generally concentrated towards the top of the breccia suggesting that the overlying dolostone has not been "mixed" extensively with the gypsum breccia. Large, laterally continuous slabs and "bedded breccia textures" in lithofacies 6 also suggest there has been little vertical movement. However, in the more chaotic gypsum breccias (section PY) the above textures are not so readily discernible. There is a

higher degree of incorporation of the Slave Point lithologies into the more chaotic exposures of lithofacies 6 than is seen in the less chaotic exposures. The fact that there is a varied degree of brecciation is substantiated by the different textures mentioned. The different breccia textures are the result of different degrees of collapse that affected lithofacies 1, and also formed lithofacies 6.

LITHOFACIES 7 (BRECCIA I)

Lithofacies 7 is a semirecessive to semiprominant particulate rubble floatbreccia that contains a medium brown to dark brown matrix and white to whitish buff clasts (Figs. 27E, 28A). Clasts are angular, extremely porous, soft, generally less than 1 cm along the longest axis and represent 30 to 40% of the lithofacies. The clasts are generally comprised of 80 to 90%, 8 micron wide calcite and 10% clear dolomite that fringes the calcite edges. This dolomite may be representative of a limpid dolomite. Clasts are either featureless or contain argillaceous laminae (Fig. 28B).

The matrix is formed of anhedral to euhedral, generally subhedral 33 micron wide calcite and up to 20 to 25%, 33 micron wide dolomite and dolomicrite. Rarely, the breccia can be completely dolomitized resulting in two bimodal dolomite crystal sizes that are both less than 4 microns and form a dense, clotted, dolomicrite. Bright red to bright orangey brown clay cutans and organic opaques occur up to

FIGURE 28

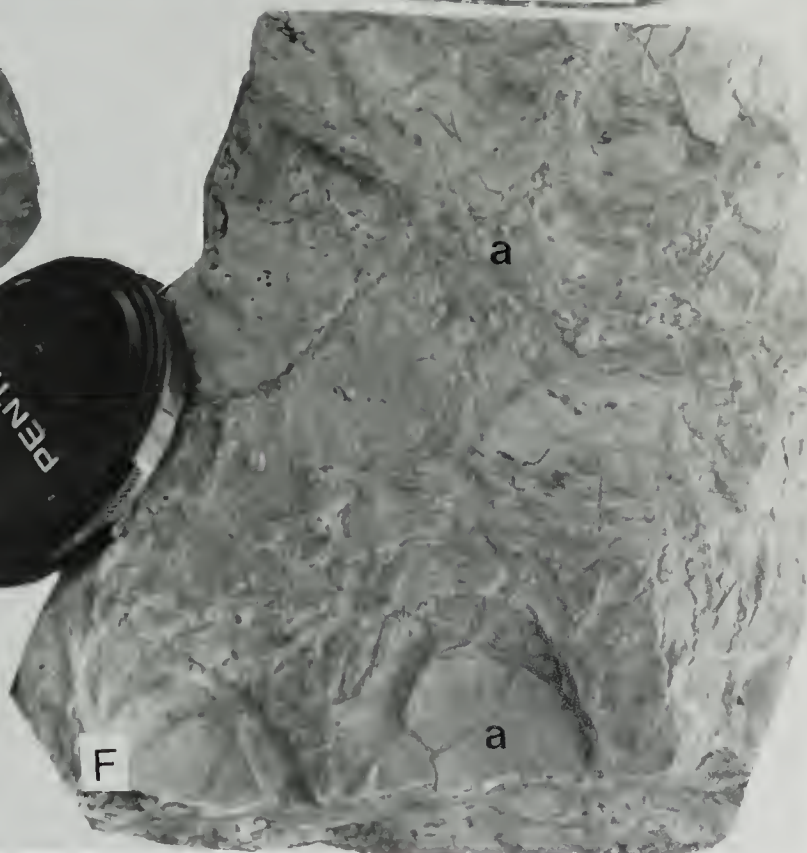
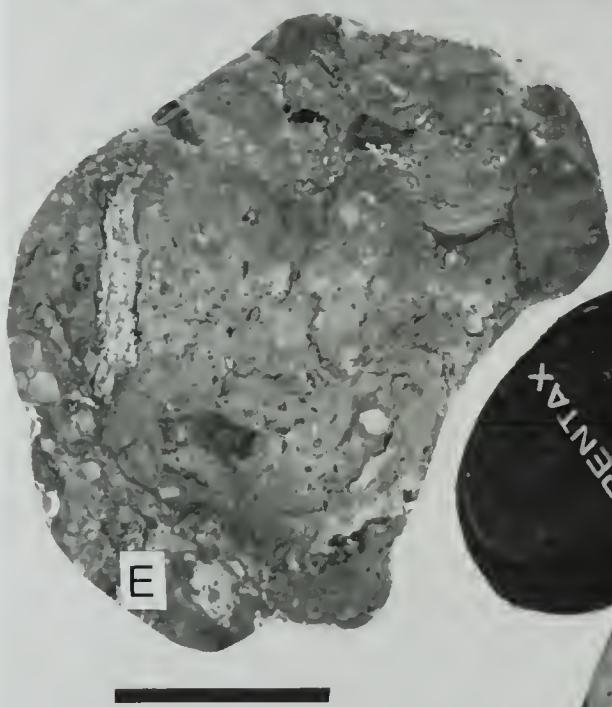
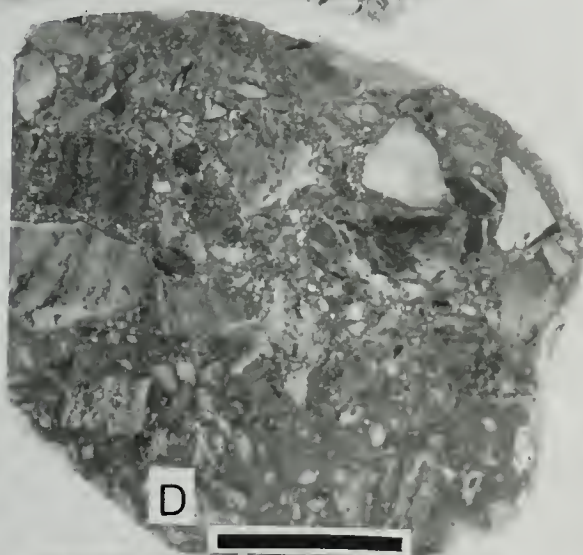
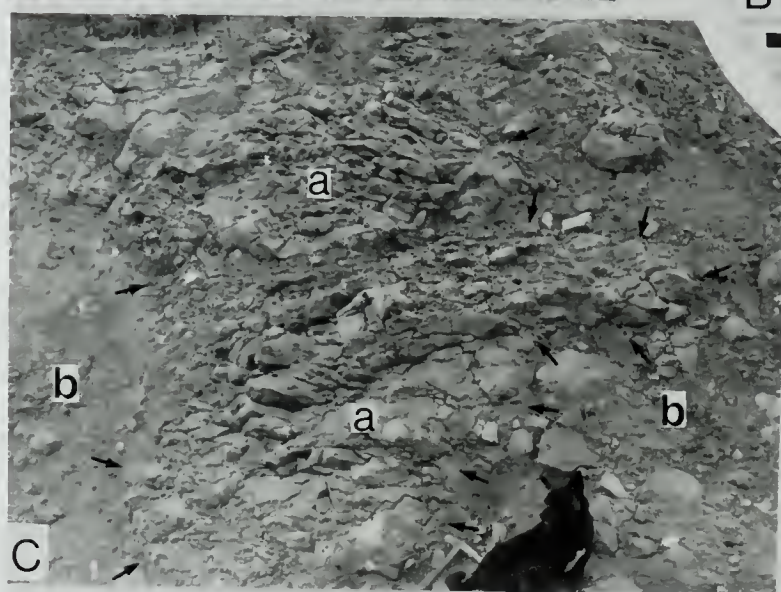
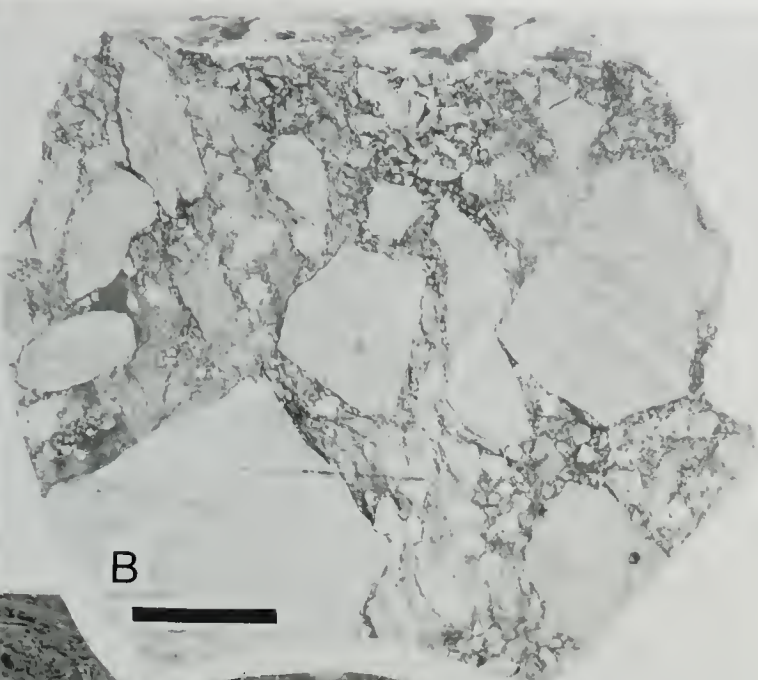
28A) Field photograph of lithofacies 7 (section PG6) showing white clasts of dolostone and dark brown, highly argillaceous limestone matrix.

28B) Hand specimen of lithofacies 7 from section PG6. Note the majority of the clasts (glaebules) are laminated. Scale bar equals 2 cm.

28C) Field photograph of lithofacies 8 (section PQ) showing high clast to matrix rock (a) and low clast to matrix rock (b). High clast to matrix rock are blocks of breccia that have not been as "mixed" as the low clast to matrix rock. A majority of the clasts are comprised of lithofacies 2c (mixing variety). The small arrows point to the edge of the high clast to matrix rocks. Hammer for scale.

28D & E) Hand specimens of matrix of lithofacies 8 showing typical particulate rubble packbreccia texture. Scale bar equals 2 cm.

28F) Hand specimen of a lithofacies 2b clast from lithofacies 8 (section PR). Clast extremely vuggy and displays mudcracks (a) and intraformational breccia on the surface. The rock is representative of supratidal conditions. Lens cap for scale.



15%, with the cutans more common in the more extensively dolomitized units.

DISTRIBUTION AND THICKNESS

Lithofacies 7 occurs as patchy, small exposures throughout the area that only reach any lateral and vertical extent in section PT, where the lithofacies is 14 m and 1.85 m, respectively.

The lower contact is gradational with the underlying bedded lithofacies 2b. The upper contact with lithofacies 8 is defined by the coarser and less homogeneous clasts size in lithofacies 8, as compared to lithofacies 7. In section PH, lithofacies 2b changes laterally into lithofacies 7, thus suggesting that lithofacies 7 is a direct chemical alteration of lithofacies 2b. In sections PK and PL, mixing of limestone and dolostone of lithofacies 2c has also chemically produced lithofacies 7 (Figs. 24D, 24E). In section PG6, breccia of lithofacies 7 (Fig. 23C) forms a distinct band that dips down into the center of a collapse area that is comprised of lithofacies 10. Overlying this breccia is a breccia band comprised solely of the white, soft, porous limestone/dolostone clasts. Towards the center of the collapse the white material becomes more massive, displaying a "zig-zag" fault pattern (Fig. 23B). Overlying this white limestone/dolostone is a breccia comprised of sandstone. This white material may be similar to the chalky

horizons of calcrete profiles (Esteban and Klappa, 1983) whereas lithofacies 7 may be a terra rossa.

TERRA ROSSA ORIGIN FOR LITHOFACIES 7

Lithofacies 7 may be representative of terra rossa. The term terra rossa has been used differently by numerous authors, but in this study refers to "...weathering products that are derived from limestone and that can be parent materials for the formation of different kinds of soils, e.g. red mediterranean, reddish brown, and other soils" (Glazovskaya and Parfenova, 1974, p. 58). Some authors (Ballagh and Rundge, 1970; Olsen et al., 1980) suggest that some of the terra rossa is not derived from the underlying limestone but from erosion of higher lying clastic sedimentary rocks. Muchkenhausen et al. (1975) stated that terra rossa formed over limestone and dolostone, but not from them. From southern Crimea, terra rossa is formed both on the products of weathering of dolomitic limestones and on calcareous breccia (Glazovskaya and Parfenova, 1974). Based on the lower contact of lithofacies 7 and its obvious formation from chemical alteration of lithofacies 2, lithofacies 7 also represents a product of weathering derived from a limestone and dolostone. Lithofacies 8 is abruptly developed above and represents, in part, a calcrete. Calcrete overlies terra rossa in Ibiza, Spain (Esteban and Klappa, 1983).

Terra rossa is typically a red to brown, shallow, sand to clay sized, granular to nutty, 10 to 70% brown clay (near 45%), that forms up to 0.61 m in depth and has no visible horizon development (Stace, 1956). Terra rossa is generally described as consisting of calcareous clayey mudstone, however, hard crystalline limestone to soft chalky limestone is also associated with terra rossa (Atkinson, 1970). Lithofacies 7 is predominately calcite, but clay/argillaceous material can comprise up to 15%.

The clasts in lithofacies 7 may represent glaeboles. "Glaebules are three dimensional units, within the soil matrix, and usually approximately prolate to equant in shape; their morphology (especially size, shape and/or internal fabric) is incompatible with their present occurrence being within a single void in the present soil material. They are recognized as units either because of a greater concentration of more constituent and/or a difference in fabric compared with the enclosing soil material, or because they have a distinct boundary with the enclosing soil material" (Brewer and Sleeman, 1964, p. 67). Based on Brewer and Sleeman's (1964) classification of glaeboles, lithofacies 7 contains nodules and papules. The former has no internal structure whereas the latter is laminated.

The red to brown color, typical of a terra rossa is due to the presence of iron oxides and oxidation of organic matter (Stace, 1956). The oxidizing activity of nonbacteria is one of the main factors in the formation of iron hydroxides and the red coloration of terra rossa (Glazovskaya and Parfenova, 1974). These authors suggested that even under similar climatic conditions and topography the reason terra rossa only develops in certain facies is the initial presence of ferric sulphides. The oxidation of sulphides by iron and sulphur bacteria results in the formation of sulphuric acid, which aids in the dissolution of the limestone. They further suggested that the presence of organic matter is due to the abundant bacterial population. Bioerosion of lithofacies 7 may also be the result of bacteria because abundant borings occur through calcite crystals in the lithofacies. The borings produce a "wormy texture" with individual borings < 1 mm wide and 2 mm long.

Clay cutans, generally brick red to orange, appear in closed voids (Glazovskaya and Parfenova, 1974) and are thought to be either insoluble residue, (Atkinson, 1970; Norrish and Rodgers, 1956) or illuviated clay (Ballagh and Rundge, 1970).

Commonly associated with terra rossa deposits is white, chalky limestone. In places, white chalky nodules occur in a brown clay matrix (Esteban and Klappa, 1983),

as white, cm thick rims around clasts (Olsen et al., 1980) or as the white underlying limestone beneath the terra rossa. Hard crystalline to soft chalky limestone is associated with terra rossa (Atkinson, 1970; Esteban and Klappa, 1983). The nodules of lithofacies 7 can be chalky, but are generally extremely porous and comprised of limestone and dolostone in varying proportions. The chalky porous limestone/dolostone occurs as breccia pipes in section PG and forms the glaebules in lithofacies 7, producing a lithofacies that looks extremely similar to terra rossa from other parts of the world (Fig. 27F).

Terra rossa can occur in small protected pockets (Gardiner and Ryan, 1962) and in fissures and sinkholes (Glazovskaya and Parfenova, 1974; Hall, 1976; Esteban and Klappa, 1983). Chafetz (1982) described 15 m thick terra rossa deposits that occupy lows on the top of pre-Late Cretaceous collapsed limestone in southwestern New Mexico. Lithofacies 7 typically occurs as discontinuous patches in paleo-dolines comprised of lithofacies 10 and in pockets, that are the result of alteration, in lithofacies 2b and 2c (Fig. 23).

The above evidence suggests that lithofacies 7 is similar to terra rossa.

LITHOFACIES 8 (BRECCIA II)

Lithofacies 8 is a prominent, complex, breccia that consists of a great variety of clast types. Some of the

clasts represent previous described lithologies whereas others are more typical of a calcrete facies.

Previously described lithologies that form clasts in lithofacies 8 are lithofacies 2a, 2b, 2c and 2d. Clasts comprised of lithofacies 2c (massive dolostone) are the most distinctive clast type in the breccia, due to their characteristic orange weathering. Clasts of lithofacies 2d are rare and are generally in those breccias displaying a "bedded texture".

Other clasts, that include laminated and non-laminated pelletal micrite and pelletal, very finely crystalline limestone may form up to 30% in localized areas throughout lithofacies 8. Other clasts that exist in trace amounts up to 5% include sandstone, black colored limestone, limestone clasts that contain quartz laminae, micaceous siltstone (Fig. 29I) and pale bluish green mudstone to brown mudstone. The previously mentioned clasts range in color from buff to pale brown, however, greenish hues and black colors are rarely seen. The black clasts, which may represent calcrete, are formed from a previous lithofacies 2b.

There are textures and features in clasts, in lithofacies 8, that are indicative of calcrete. For example, altered clasts of lithofacies 2b, as well as aggraded portions of the breccia matrix, contain ellipsoidal clasts formed of subhedral to euhedral rhombohedral calcite (Fig. 29F). It is difficult to determine whether these clasts are physically produced clasts or products of diagenetic

FIGURE 29

29A) Hand specimen of a caliche pseudoanticline from lithofacies 8, (section PA). The dashed lines mark the shape of the deformed laminae. Note how the crests of the pseudoanticlines do not coincide. An axial fracture that is filled with a sparry calcite cement (a) extends the length of the slab. Sparry calcite also occurs in spaces developed between the laminae (b). A small, localized breccia pocket forms in the corner (c). Scale bar equals 2 cm.

29B) Photomicrograph of a speleothemic clast from lithofacies 8 (section PC). The clast comprises a drusy, vertically elongated sparry calcite cement that displays feathery intercrystalline boundaries (a). A dark brown argillaceous clay replaces the sparry cement (b). Thin section under cross-polarized light. Scale bar equals 0.5 mm.

29C) Photomicrograph of possible algal spore (from lithofacies 8) showing castellated form. Thin section stained and under cross-polarized light. Scale bar equals 0.1 mm.

29D) Photomicrograph of possible calcrete from lithofacies 8. Aggraddened calcite rhombohedrals (a) have dark micrite nuclei (b) and calcite overgrowths (c). Thin section under plane-polarized light. Scale bar equals 0.5 mm.

29E) Photomicrograph of calcrete from lithofacies 8 (section PC). The two filaments represent rhizoliths (a). Thin section stained and under plane-polarized light. Scale bar equals 0.5 mm.

29F) Photomicrograph of aggraddened clast in a calcrete clast from lithofacies 8. 12 micron calcite (a) aggraddened to an anhedral mosaic of coarser calcite (b). Large rhombohedrals are due to aggradation of calcite (b) where the 12 micron calcite is plucked away from the rest of the rock by the aggradation process (d). The small arrows depict the direction of the advancing aggradation front. As the front advances the relict patches and rhombs of the 12 micron calcite are aggraded and eventually lose the rhombohedral form (e). Thin section stained and under plane-polarized light. Scale bar equals 0.5 mm.

29G) Photomicrograph of calcrete from lithofacies 8 showing common clotted texture. The clasts (a), which are the "sugary" appearing clasts in hand specimens, possibly represent caliche glaebules. Thin section stained and under plane-polarized light. Scale bar equals 0.5 mm.

29H) Photomicrograph of a floating clast comprised of a 12 micron calcite surrounded (in two dimensions) by a coarser mosaic of anhedral calcite. Force of crystallization (aggradation) of calcite has forced the clast away from the host. Arrows show the direction of the aggradation front. Thin section is stained and under cross-polarized light. Scale bar equals 0.5 mm.

29I) Photomicrograph of sandstone particles in a limestone matrix. Sample taken from lithofacies 8 (section PC). Thin section under cross-polarized light. Scale bar equals 0.5 mm

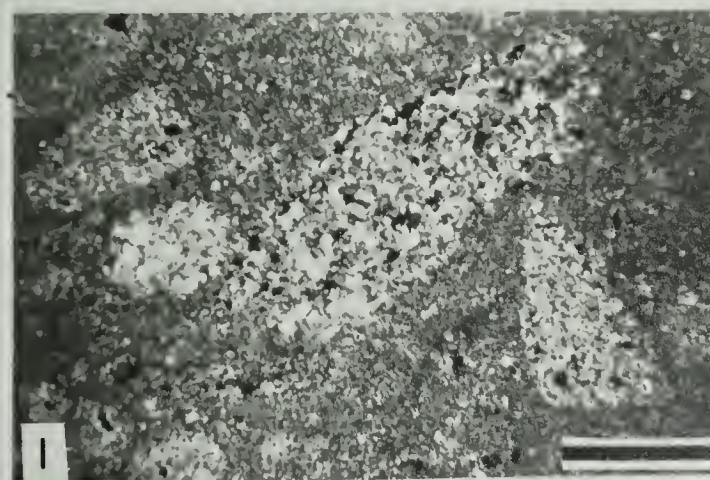
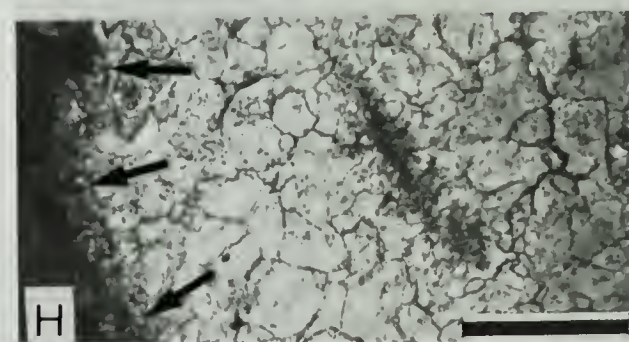
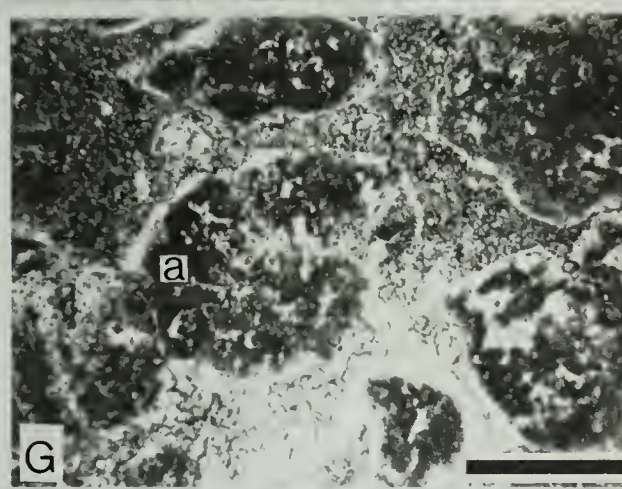
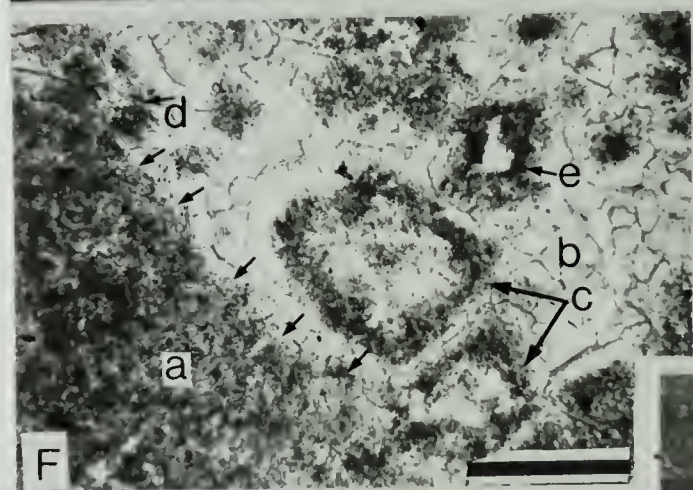
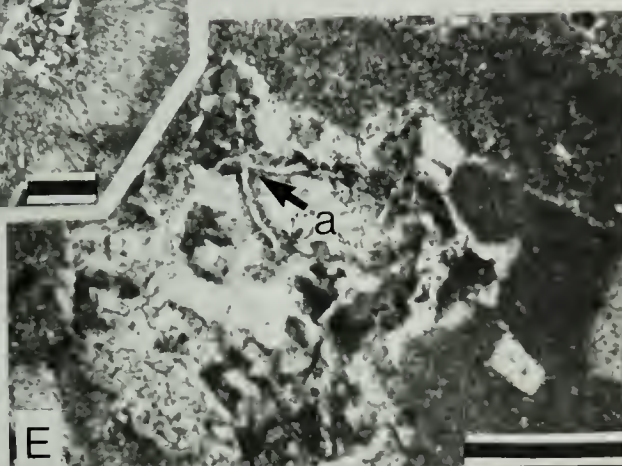
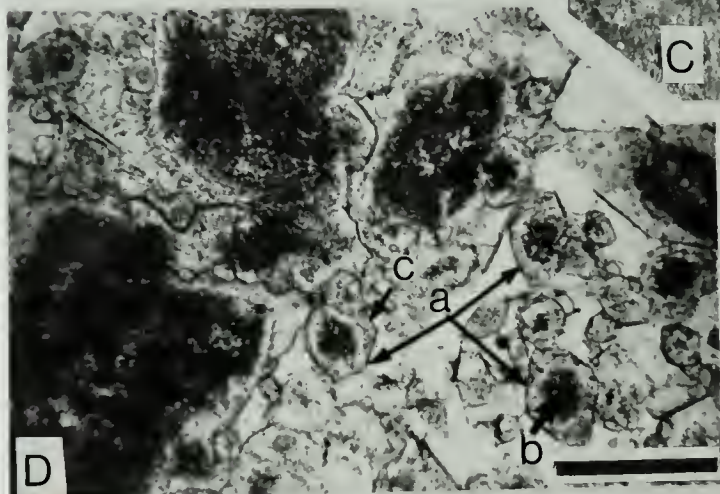
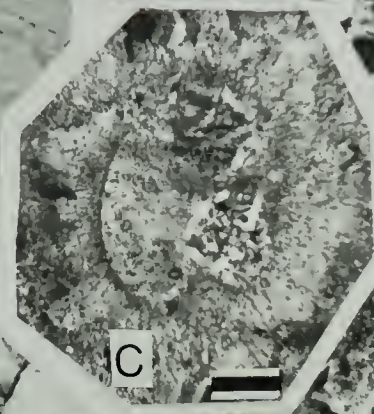
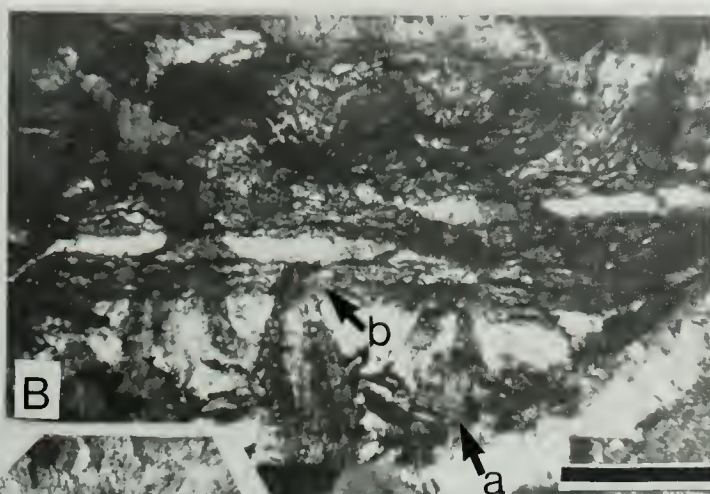
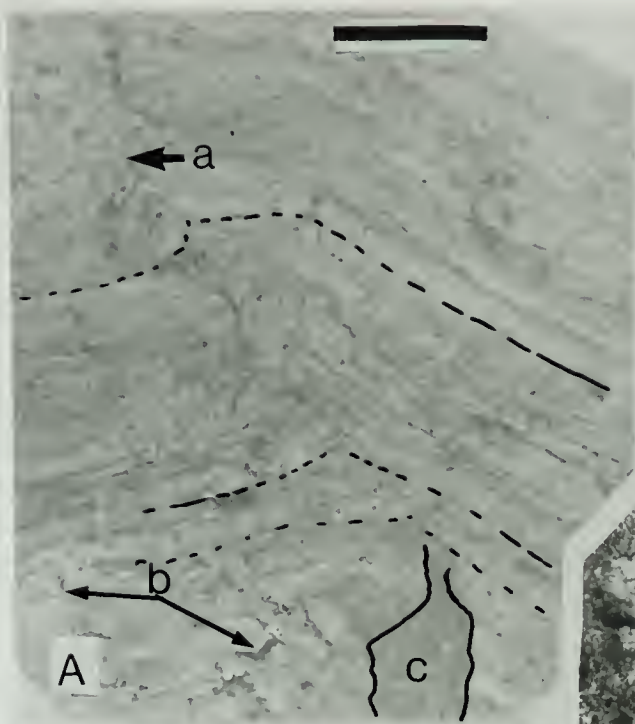


FIGURE 30

30A) Field photograph of the Slave Point Formation of section PF showing the bedded lithofacies 2b (a) underlying the brecciated lithofacies 8 (b, the two lithofacies are divided by the solid black line). Brecciated strata overlying nonbrecciated strata suggests that lithofacies 8 is a calcrete. Scale bar equals 1.5 m.

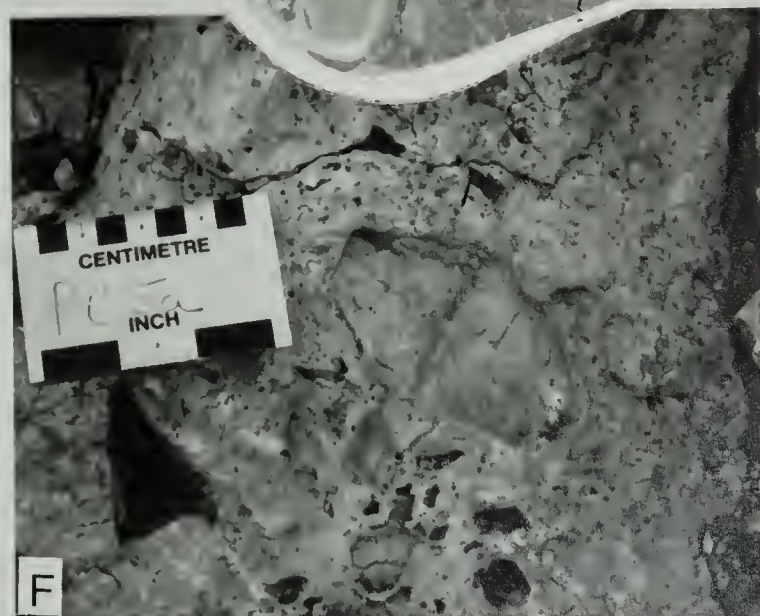
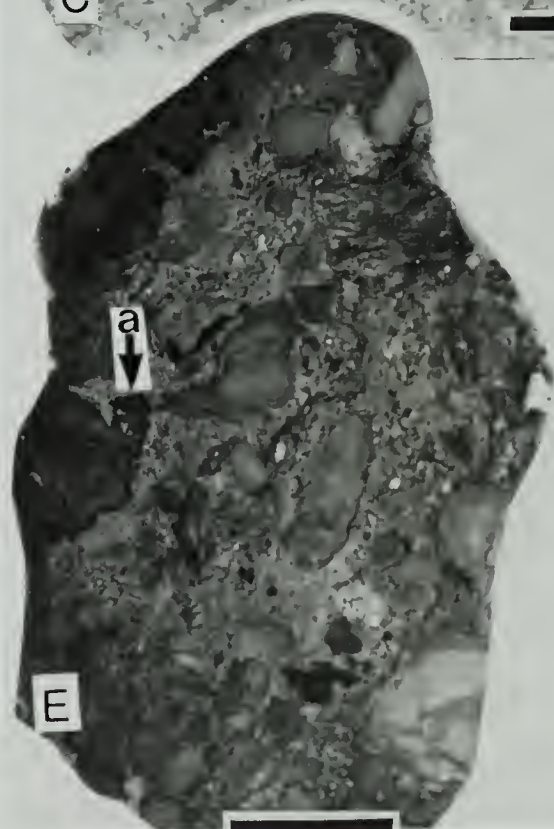
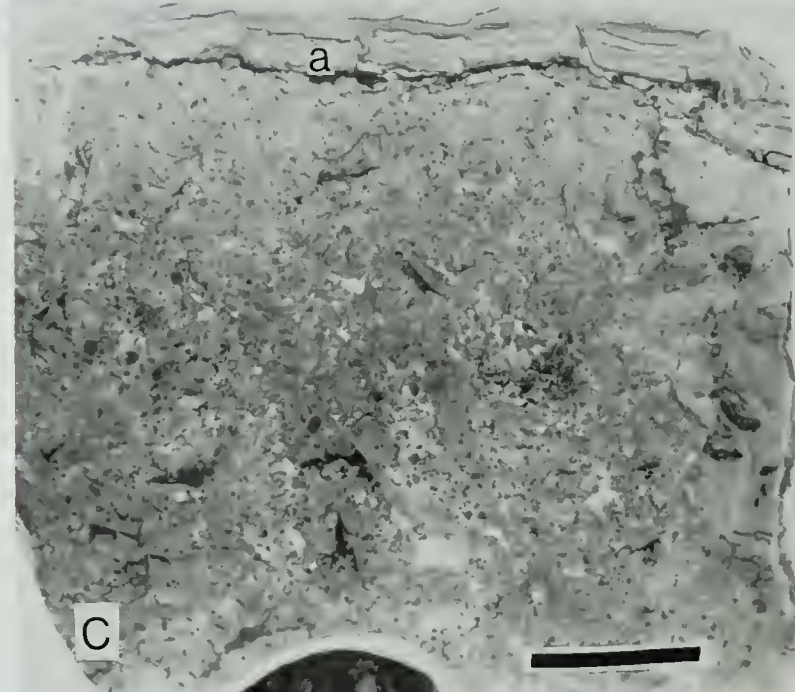
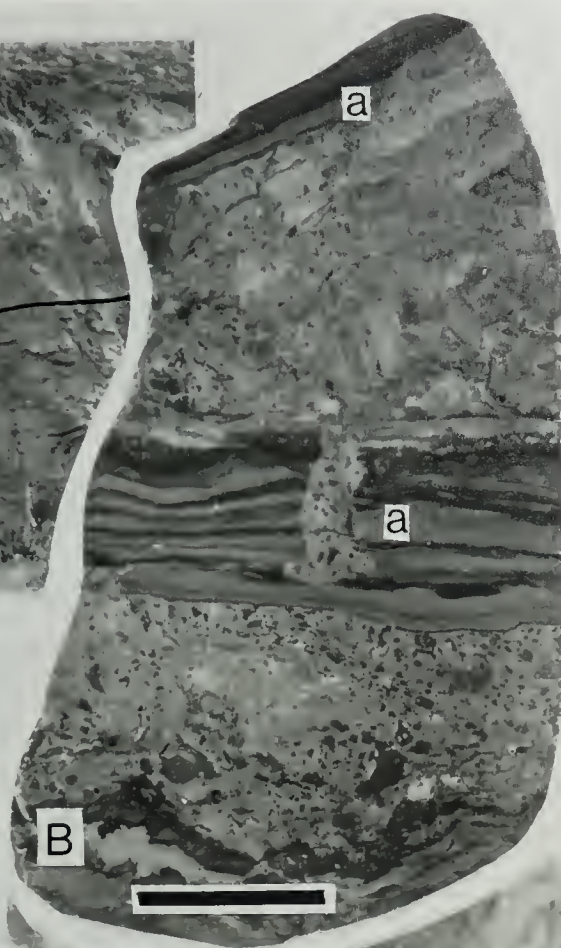
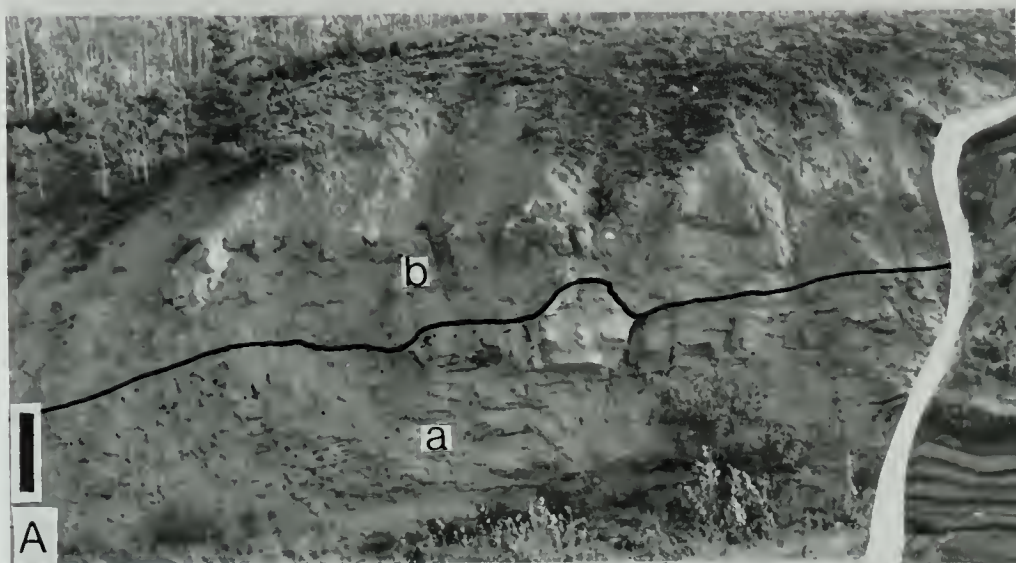
30B) Hand specimen of speleothemic clast (a) from lithofacies 8 (section PA). Scale bar equals 20 cm.

30C) Hand specimen of clast from lithofacies 8 (section PC) showing a speleothemic cap (a) near the top of the clast. Scale bar equals 2 cm.

30D) Hand specimen of pisolites from lithofacies 8 (section PA) showing well developed micrite coats (a) and thinning (b) and thickening (c) of the coats around the nucleus. The matrix is comprised of small (oolites) coated grains. Scale bar equals 2 cm

30E) Hand specimen of clast from lithofacies 8 showing black limestone clasts (a). Scale bar equals 2 cm.

30F) Field photograph of lithofacies 8 (section PC) displaying a boxwork-like texture. However, solution pits are not due to dissolution of evaporite, but due to dissolution of lithofacies 2c dolostone clasts



alteration of the matrix, or both. The rhombohedral calcite becomes larger towards the center of the clast, but the rhombs are always concentrated toward the outer edges of the clasts. The rhombs are 20 microns wide with centers comprised of 12 micron wide calcite and micrite that give the rhombs a zoned appearance (Figs. 29D, 29F). These rhombs may be representative of dedolomite. A later phase of calcite cement commonly occurs as overgrowths on the rhombs (Fig. 29D). The larger rhombohedral calcite is commonly rimmed with clear dolomite that is similar to limpid dolomite. Rarely this clear dolomite forms amorphous blebs in the calcite rhombs. These clasts, as well as portions of the matrix of the breccia commonly display a clotted, sugary texture, as well as indistinct alveolar textures (Figs. 29E, 29G).

Clasts similar to lithofacies 2a also occur in the breccia. However, these clasts are diagenetically more altered and contain pisolites which have well developed, dense, micritic coatings (Fig. 30D). The micritic coats reach a maximum thickness of 4 mm, but thinning and thickening of the micritic coats by as much as 2.5 mm is common. Faint laminations in the micritic coats are difficult to follow laterally, but the number of laminations are as few as 4 to 5. The pisolites, are a maximum of 3 cm long and 2 cm wide. The nucleus of the pisolites is generally ellipsoidal in shape with the micritic coat not always forming the same shape as the nucleus. Some coats can

be more angular, and in places almost display well developed corners (Fig. 30D). Pisolites contained in caliche from the Barbados shows no consistent preferred elongation or thickening of micrite coats (Harrison, 1977). However, in some of the pisolites from the study area both the top and the bottom of the pisolites, as well as the micritic coats, are concave downwards. Peryt (1983) stated that the cortex of some pisolites are sometimes asymmetrical, with the coatings showing downward thickening. In pisolites of lithofacies 8 aggraded calcite forms clotted clusters in the nucleus, thus giving it a composite texture. Clear, sparry calcite can form between the clotted nucleus and the micritic crust. The pisolites float in a matrix formed of 0.5 mm diameter coated grains. Clasts of this type are most abundant in the west end of section PA.

Esteban (1976) gave criteria for distinguishing vadose pisolites from caliche pisolites. Pisolites from lithofacies 8 are similar to his caliche pisolites in that the pisolites are basically micritic with a clotted peloid texture, microspar is locally important and there are rarely more than 5 laminae. However, Peryt (1983) argued that there are numerous exceptions to the number of laminae. Based on a similarity to Esteban's caliche pisolites, the smaller and larger coated grains are thought to have formed in a caliche environment and using Peryt's (1983) nomenclature are named caliche pisoids and caliche microids, respectively.

Speleothemic clasts and crusts are rarely found in lithofacies 8. Crusts are tabular in shape whereas the clasts are generally rectangular (Figs. 30C, 30B). Speleothemic clasts are formed of a drusy calcite that displays a feathery appearance and planar to highly irregular intercrystalline boundaries (Fig. 29B). This texture is similar to calcite in indurated travertine from central Texas (Chafetz and Butler, 1980, p. 512). The calcite is elongated vertically with some crystals forming radiating clusters. The speleothem clasts are generally cut and replaced in a mottled, clotted and laminated fashion by a dark brown opaque clay (Fig. 29B).

Other features in lithofacies 8, which are distinctive of calcrete (Esteban and Klappa, 1983) are pseudoanticlines, black limestone clasts (Fig. 30E), and clasts that contain bleached borders.

Paleozoic examples of calcrete or karst are rarely reported in the literature; therefore, comparison of ancient examples with Holocene and Tertiary examples is also rare. This may reflect the difficulty in detecting ancient examples. However, the clasts described above may be of a calcrete origin.

Clast sizes in lithofacies 8 range from 2 mm up to 7.5 X 3 m, with a modal clast size of 4 to 6 cm along the longest axis. In section PA, a 0.5 m thick horizontal slab extends for 20 m across lithofacies 8. This slab contains caliche-pseudoanticlines (Fig. 29A). Clast shapes are

generally square to blocky, but tabular and rectangular clasts are common. The larger clasts are subangular to angular and are generally more angular than the 1 to 5 cm clasts. The clast to matrix ratio ranges from 80:20 to 30:70 with an average of approximately 40:60.

The matrix of lithofacies 8 is predominantly medium to dark brown limestone that consists of three modal calcite crystal sizes; namely, > 4 micron wide calcite, a 12 micron wide calcite and a > 20 micron wide calcite.

Micrite occurs as clotted clusters or peloids and may form up to 30% of some samples. Micrite is commonly found in rhombohedral calcite, (as described earlier) and associated with an alveolar-like texture. In places, it is hard to differentiate between relict micrite and possible micritization.

The second modal calcite crystal size ranges from 7 to 15 microns in width with a 12 micron mode. It forms as an anhedral mosaic ranging from 40 to 85% of the matrix. In many areas it occurs with a 12 micron wide dolomite where an inverse size relationship exists between the two crystal types. When the dolomite is slightly larger than 12 microns the calcite is smaller than the 12 micron wide mode.

The third calcite type is rhombohedral calcite > 20 microns wide. This calcite forms anhedral to subhedral crystals when it occurs in clusters, subhedral to euhedral crystals when isolated and as channels throughout the matrix. Intercrystalline boundaries are slightly curved and

sutured, however, planar boundaries do occur. The rhombohedrals are dirty and have a high inclusion content. Water clear dolomite fringes the outer edges of the rhombohedrals and forms a meniscus cement between rhombohedrals. These rhombohedrals form diagenetic clusters that eventually reach clast sizes (as described earlier). Micrite and 12 micron diameter calcite occur in this calcite type.

Large, clear, sparry calcite, with planar intercrystalline boundaries, occurs rarely in fractures in lithofacies 8. Structures similar to algal spores occur in the matrix (Fig. 29C). This breccia is a particulate rubble floatbreccia.

In lithofacies 8 there are areas that are more extensively brecciated than others. In the "more bedded" breccia, large areas of the rock have a higher matrix to clast ratio than the surrounding breccia. For example, in section PQ, blocks of "less brecciated" breccia are surrounded by a breccia that is more chaotic and has a noticeably higher matrix to clast ratio (Fig. 28C). The rock with the higher matrix to clast ratio contains more green shale than the low matrix to clast ratio rock. In the low matrix to clast ratio rock the shale content is in the 15 to 25% range. In places lithofacies 8 becomes brecciated, but not completely churned, because beds can be followed for relatively short distances. However, where vertical movement is greater, the breccia is more chaotic and the Peace Point

Shale gets incorporated into the breccia more readily. Those areas where shaly, high matrix to clast type breccia surrounds less chaotically mixed breccia, represent either the edge of a major collapse or an area where lithofacies 8 has not undergone a severe degree of brecciation.

DISTRIBUTION AND THICKNESS

Lithofacies 8 reaches a maximum thickness of 17 m in section PB. This lithofacies, when compared to the other lithofacies, is the most laterally extensive in the study area. It occurs in sections PA, PH, PK, PQ, PR, PT, PV, PX and PZ. A maximum lateral extent of 380 m occurs in sections PA, PB and PC. Lithofacies 8 thickens in lows developed between; 1) the lithofacies 1 anticlines (section PX), 2) the domal shaped, laterally continuous outcrops of lithofacies 6 (sections PA and PD), 3) lithofacies 6 and lithofacies 1. In the lows just mentioned and on top of lithofacies 6, lithofacies 8 displays a chaotic texture. However, where the breccia occurs on top of non-brecciated, slightly deformed sequences, such as lithofacies 2b (east end of section PE, section PZ, and the west end of section PR) the breccia is less chaotic and indistinct bedding occurs in the breccia. The occurrence of lithofacies 8 above non-brecciated strata suggests that the breccia is not a result of collapse, because the underlying unit is not brecciated (Fig. 30A). This fact along with the associated textures of lithofacies 8 substantiates the

idea that portions of lithofacies 8 are calcrete.

The lower contact of lithofacies 8 is generally gradational with the underlying lithofacies 6, but a gradational contact also occurs with lithofacies 7 (east end of section PT, west end of section PV and section PK). The lithofacies 7/lithofacies 8 contact always occurs on the edge of lithofacies 1. Generally, at those localities a gradational vertical sequence forms, in ascending order; lithofacies 2, lithofacies 7, and lithofacies 8.

The upper contact of lithofacies 8 occurs only in section PK where lithofacies 9 overlies with an indistinct, gradational contact.

LITHOFACIES 9 (BRECCIA III)

Lithofacies 9 is a semirecessive to semiprominent, buff to pale light brown to medium brown, matrix supported, particulate rubble floatbreccia, that has an indistinct bedded breccia texture (Fig. 31A). The lithofacies fines upward with clasts and shale content decreasing and increasing, respectively, towards the top of the exposure. The clast to matrix ratio decreases from 80:20 at the base to 40:60 near the top.

Clast lithologies are nearly 100% composed of limestone, which include massive and laminated micrite. These limestones are variably dolomitic and display a granular, silty texture, different from the limestone seen in the previous lithologies. However, limestone typical of

FIGURE 31

31A) Field photograph of east end of section PL with arrows pointing to lithofacies 9 (a)/lithofacies 8 (b) contact. Note the more recessive nature of lithofacies 9, due to a higher shale content than lithofacies 8. Scale bar equals 1.5 m

31B) Hand specimen of matrix of lithofacies 9 from section PL. Note the higher percentage of clay sized material in the matrix than seen in lithofacies 8. Scale bar equals 2 cm.

31C) Hand specimen of lithofacies 9 (section PG4) showing matrix of lithofacies 9. Note the green shale particles (pale color, a) and the dark and light brown shale particles (b) Scale bar equals 2 cm

31D) Hand specimen of matrix of lithofacies 9 (section PC) from breccia pipe in lithofacies 8. Scale bar equals 2 cm

31E) Field photograph of lithofacies 10 (section PG2). Heavy arrow points to the contact between lithofacies 8 (calcrete portion, a) and lithofacies 2b (near the base) and lithofacies 10 (b). Sandstone blocks (sst) are the largest block type in this breccia. A small sandstone fissure extends into lithofacies 2b (c). Scale bar equals 1.5 m.

31F) Hand specimen of calcareous sandstone block from lithofacies 10. Note the pitted texture of the sample. Scale bar equals 2 cm.

31G) Field photograph of black, carbonaceous shaly matrix of lithofacies 10 (section PG2) showing black carbonaceous shale (a) and carbonaceous siltstone/very fine grained sandstone clasts (b). Lens cap for scale (arrow).

31H) Field photograph of lithofacies 10 (section PG2) showing bedded calcareous sandstone on extreme eastern edge of collapse. Hammer for scale.

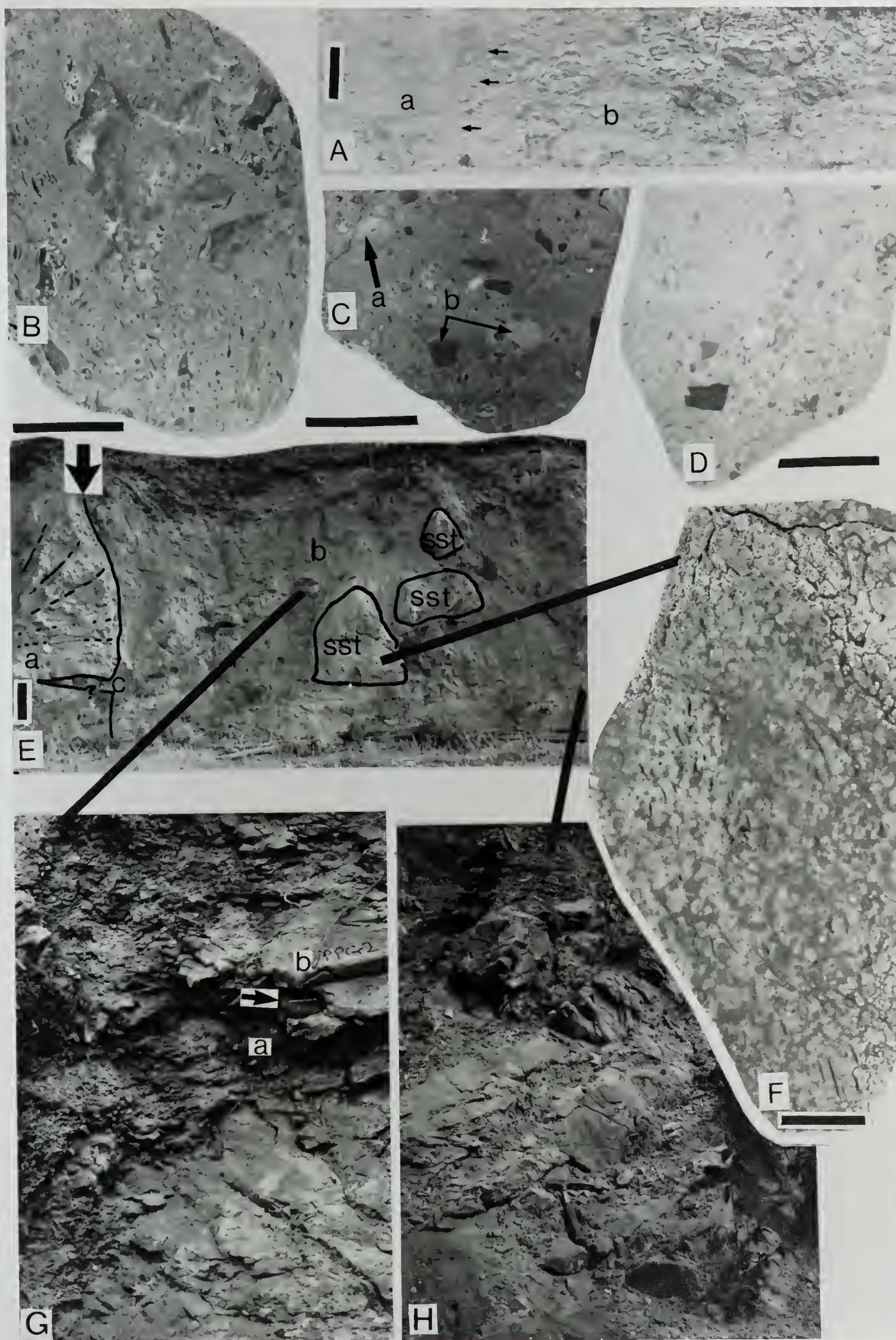


FIGURE 32

32A) Hand specimen of calcareous sandstone from a doline in the Cretaceous outlier of central Wood Buffalo National Park. This sandstone, with the exception of sedimentary bedding, is identical to calcareous sandstone found in breccias in the Peace Point Cliffs. Scale bar equals 2 cm.

32B) Hand specimen of carbonaceous very fine grained sandstone/siltstone (section PG2). Note the burrow/soft sediment deformation (?) structures (a). Scale bar equals 2 cm.

32C) Hand specimen of calcareous sandstone (section PG2). Dark patches are carbonaceous and shaly. Note that there is no pitted texture. Scale bar equals 2 cm.

32D) Hand specimen of carbonaceous siltstone/very fine grained sandstone (section PG6m) showing carbonaceous streaks (dark areas) and "burrow-like" structures (a). Scale bar equals 2 cm.

32E) Photomicrograph of siltstone/very fine grained sandstone showing the high argillaceous content of the matrix (dark area). Thin section under plane-polarized light. Scale bar equals 0.5 mm.

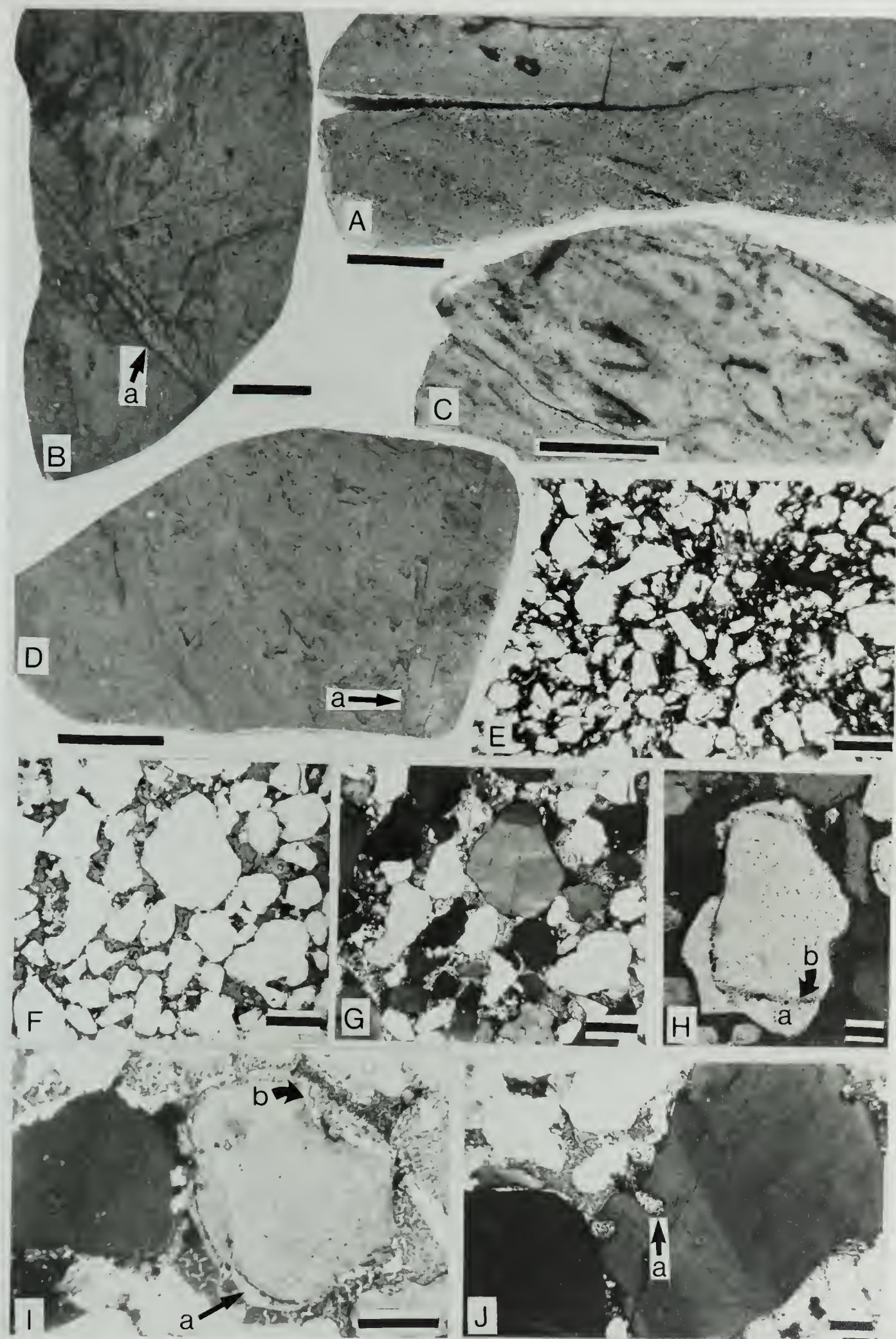
32F) Photomicrograph of calcareous sandstone (section PG6) showing quartz grains (white areas) and a matrix comprised of sparry calcite cement (dark areas). Thin section stained and under plane-polarized light. Scale bar equals 0.5 mm.

32G) Photomicrograph same as (F), only under cross-polarized light. Note the undulatory extinction of the central quartz grain. Scale bar equals 0.5 mm.

32H) Photomicrograph of calcareous sandstone (section PG6) showing well developed quartz overgrowths (a) that are distinguished from the original quartz grain by a "dusty" rim (b). Thin section stained and under cross-polarized light. Scale bar equals 0.1 mm.

32I) Photomicrograph of calcareous sandstone showing quartz overgrowth (a) on a quartz grain. Note how the calcite corrodes the overgrowth and the grain (b). Thin section stained and under plane-polarized light. Scale bar equals 0.1 mm.

32J) Photomicrograph of calcareous sandstone showing the corroding affect of the calcite with microembayments (a) of calcite into the quartz. Thin section stained and under plane-polarized light. Scale bar equals 0.1 mm.



lithofacies 2c occurs near the base of this lithofacies. The clasts are angular to subangular and are only up to 1 m long, along the longest axis.

The matrix comprises 49 to 53% of 10 micron wide anhedral calcite, 10 to 14% of 10 micron wide anhedral dolomite and 10% opaques and argillaceous material. The matrix also contains particles under 1 cm that comprise up to 13% brown shale, 12% pale green shale and 5% variably dolomitic limestone clasts (Figs. 19B, 19C, 19D). The higher amount of shale in this lithofacies, as compared to the other lithofacies, is reflected in these small particles that form part of the matrix and in shale that is mixed in with the matrix, thereby resulting in the green color.

DISTRIBUTION AND THICKNESS

Lithofacies 9 is not a common lithofacies and occurs in a 120 m long and approximately 5 m high exposure. The lower contact is gradational with lithofacies 8, based on the mixing of lithofacies 8 and 9 at the top of lithofacies 8. The upper contact is not exposed in the study area.

LITHOFACIES 10 (SANDSTONE BRECCIA)

Lithofacies 10 is the most complex and varied breccia in the study area. Clasts are representative of almost all of the lithofacies including sandstone clasts of Cretaceous age. The clasts are semirecessive to semiprominent, whereas the matrix is recessive (Fig. 31E). The sandstone breccia has an overall greyish weathering color due to the color of

the weathered matrix.

Lithofacies 10 contains two distinct sandstone lithologies (Figs. 31F, 32B). The first sandstone is a light to medium grey, orangey buff to medium grey weathering, grain supported, sandstone that has a grain to matrix ratio generally ranging from 70:30 to 65:35. Grain sizes range from 0.06 mm to 3.5 mm with a medium grain size mode. The grains are predominantly quartz, which typically forms 48 to 55% of the rock. The quartz is moderately clean, moderately strained and contains well developed quartz overgrowths (Figs. 32G, 32H). The border of the original grain is outlined by rims of argillaceous material and inclusions. Polycrystalline, metamorphic quartz, displaying strongly sutured intercrystalline boundaries, is generally 2 to 4% of the rock. Other minerals include up to 5% chert and 2 to 3% calcic feldspar, microcline and coal. Brown and green shales, ranging from trace amounts up to 7%, are squeezed around the more competent grains.

The grains are angular to well rounded, but generally well rounded. Corrosion of the quartz by calcite cement causes the grains to appear more angular than they originally were. The grains display poor to moderate sorting and a low to high sphericity with a moderate sphericity mode.

The matrix, which forms 8 to 15% of the sandstone, comprises up to 10% organic and argillaceous matter, and < 5% silt sized quartz, muscovite and bitumen. Intergranular

porosity is up to 25% in some samples.

The sandstone is cemented with a calcite cement which consists of single, dirty, large crystals generally 100 to 200 microns wide.

The sandstone displays a variety of textures. Corrosion of the quartz grains and quartz overgrowths by calcite cement, produces micro embayments in the grains (Figs. 32I, 32J) . Minor pressure solution and concavo-convex contacts also occur. Where slabbed, the rock has a pitted texture, probably due to leaching of the calcite cement or localized loss of the more argillaceous/organic patches (Fig. 31F). Other sandstones, which are lithologically identical do not display a "pitted surface". These sandstones rarely contained grains which reached pebble sized proportions. Rarely, there is a bimodal grain distribution, with 40% of the grains very fine grained and 20% medium grained. The larger grain size is more rounded and has a better developed sphericity than the smaller grain size. Both sandstones are generally featureless, however some do display rare carbonaceous, wispy, laminations and contorted laminae (Fig. 32C). The sandstone is best named a medium grained calcareous sublitharenite.

The second type of sandstone comprises a medium to dark grey, orangey brown, oxidized, yellow sulphur stained, matrix supported, siltstone to very fine grained sandstone with a grain to matrix ratio of 45:55 (Figs. 32B, 32D). Grain sizes range from silt sized to coarse grained, with a

modal grain size of silt to very fine grained. Grains comprise 45 to 50% quartz, < 5% shale, muscovite flakes, metamorphic quartz, chert, calcic feldspars and microcline.

Grains are angular to subrounded with a subangular mode and display a low sphericity. Sorting is moderate in the siltstone and poor in the fine grained sandstones. Quartz overgrowths commonly occur on the larger grains, with minor pressure solution between quartz grains.

In the sandstone, the matrix consists of 25 to 35% argillaceous and carbonaceous/organic material, up to 20% silt sized quartz and chert, and up to 5% silt sized muscovite flakes and shale (Fig. 32E).

Clasts of the sandstone contain abundant carbonaceous, coalified, organic, wispy, partings, laterally discontinuous laminations, rare to common microfaults and indistinct, vertical burrows (Figs. 32B, 32D). The clasts are highly contorted, displaying a soft sediment type of texture. This sandstone ranges from a very fine grained immature sublitharenite to a very fine grained immature quartzarenite.

Other clasts in lithofacies 10, in ascending order of abundance, are formed of lithofacies 2c, 2b, and 7.

Clast shapes are difficult to discern; however, they are generally rounded and elongated along one axis. Sandstone blocks are up to 13.5 m long and 4.5 m high. The larger blocks are formed of medium grained sandstone, whereas the smaller clasts are formed of the carbonaceous,

very fine grained sandstone/siltstone (Figs. 31E, 31G). Some blocks (section PG5) are large, bedded, composite structures, that can attain lengths of 45 m, also displaying well developed vertical sequences. For example, section PG5 is a bedded block that is comprised of a basal medium grained sandstone, overlain by lithofacies 2b, which is overlain by lithofacies 7.

Large white breccia pipes are commonly associated with the sandstone breccias in section PG. Throughout section PG these pipes commonly display wide bases but taper towards the top of the pipe, thus defining a shape like an inverted cone. These pipes weather white and contain up to 95% chalky, extremely permeable, soft, variably calcareous, dolostone. Clasts of lithofacies 2c (the mixing variety) and lithofacies 2d form up to 5% of the pipes. In section PG6 the pipes reach 15 m in height with clast sizes no greater than 1 m, and generally averaging 20 cm. These pipes commonly occur on the edges of collapses containing lithofacies 10. There are no sandstone clasts in these pipes, thereby suggesting that they formed prior to sandstone deposition.

The block to matrix ratio of lithofacies 10 is hard to determine because the matrix is recessive and hard to separate from talus near the base of the cliffs. However, the matrix is a light grey, greenish grey, medium brown to black, rusty orange to yellow, extremely friable and contorted shale/mudstone (Fig. 31G). It also contains

nodular iron/siderite concretions. Clasts of medium grained sandstone and carbonaceous, very fine grained sandstone/siltstone commonly form, laterally discontinuous, interbedded units with the shale. However, beds can only be traced laterally for 1 m (section PG6). In section PG6, white, 7 mm wide clasts of lithofacies 7 (Fig. 23C) and 20 cm long, medium grained, sandstone lenses occur in the carbonaceous shaly matrix. The shaly matrix generally occurs at the base of the cliffs and becomes more silty towards the top of the cliff.

DISTRIBUTION AND THICKNESS

Lithofacies 10 occurs in sections PG1 to PG6. In section PG6 the lithofacies is approximately 13 m thick. Section PG (PG1 to PG6), approximately 180 m in length, is comprised of extensively brecciated zones (lithofacies 10) that contain large blocks of sandstone and laterally adjacent areas that contain lithofacies 2, 7, and 8, but no sandstone. In section PG, lithofacies 10 has been collapsed down adjacent to the older strata of the Slave Point Formation (Fig. 31E).

Lithofacies 10 occurs as locally restricted breccias that are laterally defined by the distinct bending down of beds adjacent to the other mentioned lithofacies. For example, on the edge of lithofacies 10, in section PG3, lithofacies 2d and 2c abruptly bend down into the brecciated zone. Ropey textures, typical of lithofacies 2d, as well as small folds, occur on the

edge of the collapse (Figs. 23E, 23F). In section PG6 medium grained sandstone beds of lithofacies 10 also dip toward the center of collapse (Fig. 31H). In section PG1, on the extreme western edge of lithofacies 10, fissures that contain the medium grained sandstone (of lithofacies 10) extend laterally into lithofacies 2b of section PG1 (Fig. 31E). The sandstone in the fissure is capped by a 0.2 m thick band of laterally continuous laminated greyish green shale.

Throughout section PG, lithofacies 10 is adjacent to bedded to slightly brecciated vertical sequences of lithofacies 2, 7 and 8. The contact between lithofacies 10 and lithofacies 2, 7, and 8 is not sharp, but the prominent nature of lithofacies 2, 7 and 8 are in direct contrast to the recessive nature of the sandstone breccia. On the eastern edge of section PG the shaly matrix of lithofacies 10 thins to a wedge between lithofacies 2b and 2c of section PH (Fig. 21C).

Lithofacies 10 appears to cut vertically through the older lithofacies 2, 7, and 8.

CRETACEOUS STRATA IN BRECCIAS AT PEACE POINT

Calcareous sandstone, carbonaceous siltstone to very fine grained sandstone and black shale/mudstone are contained in breccias representing lithofacies 10. Palynology samples taken from this lithofacies were analyzed by Singh (Appendix 5). The flora reported consistently placed an Upper Barremian to Aptian age on the Early

Cretaceous strata, however, some floras were deposited during the Santonian to Maestrichtian (Late Cretaceous). Based on the flora present, Singh (Appendix 5) has equated the calcareous sandstone, carbonaceous siltstone to very fine grained sandstone and black shale/mudstone to the McMurray Formation. Lithologically, the calcareous sandstone at Peace Point is identical to calcareous sandstone sampled from a doline in the Cretaceous outlier of Wood Buffalo National Park. Flora, indicative of the McMurray Formation has been described from both areas (Singh, Appendix 5).

Recent forms were rarely seen in samples from lithofacies 10, however, the presence of these forms can be explained by contamination by surficial debris.

Some samples collected contained only Devonian spores, whereas others contained a mixture of Devonian and Cretaceous floras. A mixture of Devonian and Cretaceous floras may be the result of mixture during collapse or a mixing during deposition. The latter suggestion is the more likely because as the Cretaceous seas transgressed over the Wood Buffalo National Park area, Devonian floras were incorporated into the Cretaceous sediment. Therefore, those samples with both Devonian and Cretaceous flora may be representative of the Devonian-Cretaceous contact. From the samples analyzed the consistently oldest Cretaceous age for the sandstones, siltstones and shales is Upper Barremian, thereby suggesting that these were the first sediments deposited after the Post-Waterways (Peace Point Member)

unconformity.

B. GRADATION BETWEEN LITHOFACIES

INTRODUCTION

There are 3 main episodes of collapse represented by strata of the Gypsum Cliffs area. Collapse 1 is a minor collapse that occurred prior to deposition of the shales of the Peace Point Member and prior to the severe diagenesis that altered the original lithology of the Slave Point Formation. Collapse 2 is the major breccia forming event and is responsible for the gradation among the majority of the lithofacies. Collapse 3 occurred after deposition of the Cretaceous sandstone but prior to glaciation.

The formation of the breccias (lithofacies 6, 8 and 9) represented a single brecciation event that is represented by collapse 2. Lithofacies 7 and the calcrete portion of lithofacies 8, however, formed earlier due to subaerial exposure; but were later brecciated and incorporated into lithofacies 8 during collapse 2. Each breccia is gradational with respect to the degree of brecciation, as well as being gradational to the breccias stratigraphically above and below. Gradation among the lithofacies with respect to contacts, shale content and stratigraphic position suggests that collapse 2 was a single, major event.

Selected areas of this study have been chosen so that the gradation among the lithofacies can be highlighted. For example, sections PK and PT are typical of the transition of

bedded strata to breccia that is typically developed on the edges of major collapsed zones. There are numerous localities where sequences and phenomena typical of sections PK and PT occur (sections PV, PX and PZ); however, sections PK and PT should provide an insight as to why the brecciation process is thought to be gradational and why the second phase of collapse observed at Peace Point is considered to be a single event.

GRADATION AMONG LITHOFACIES 2, 7, 8 AND 9

SECTION PK

A well exposed outcrop displaying the gradation between lithofacies 2, 7, 8 and 9 occurs in section PK (Fig. 33A).

In section PK, the Slave Point Formation is highly contorted, but is bedded parallel to the erosive contact that truncates the underlying lithofacies 1. Working upsection, from the top of lithofacies 1, 0.6 m of lithofacies 2b is the first bedded unit developed on top of the unconformity. An overlying 0.9 m thick unit is comprised of lithofacies 2b, 2c and 2d with a massive, basically pure dolostone concentrated toward the top of the outcrop (Figs. 33B, 33C). All of the units mentioned above become progressively more brecciated westward, where they gradationally change to lithofacies 8 (Fig. 33D).

Near the top of the outcrop a large isolated, block of lithofacies 2c (mixing variety) occurs (Figs. 33A,

FIGURE 33

33A) Field photograph of section PK showing the Fort Vermilion Formation/ Slave Point Formation contact (solid black line). Note the scalloped nature of the contact and the truncation of the Fort Vermilion Formation (a). Eight meters of lithofacies 1 (Fort Vermilion Formation) strata are missing at the contact (at arrow). Bedded to slightly brecciated strata of the slave Point Formation (b) overlie the Fort Vermilion strata. The outcrop is on the extreme edge of a major collapse where the center of the collapse is to the left of the photo. Scale bar equals 1.5 m.

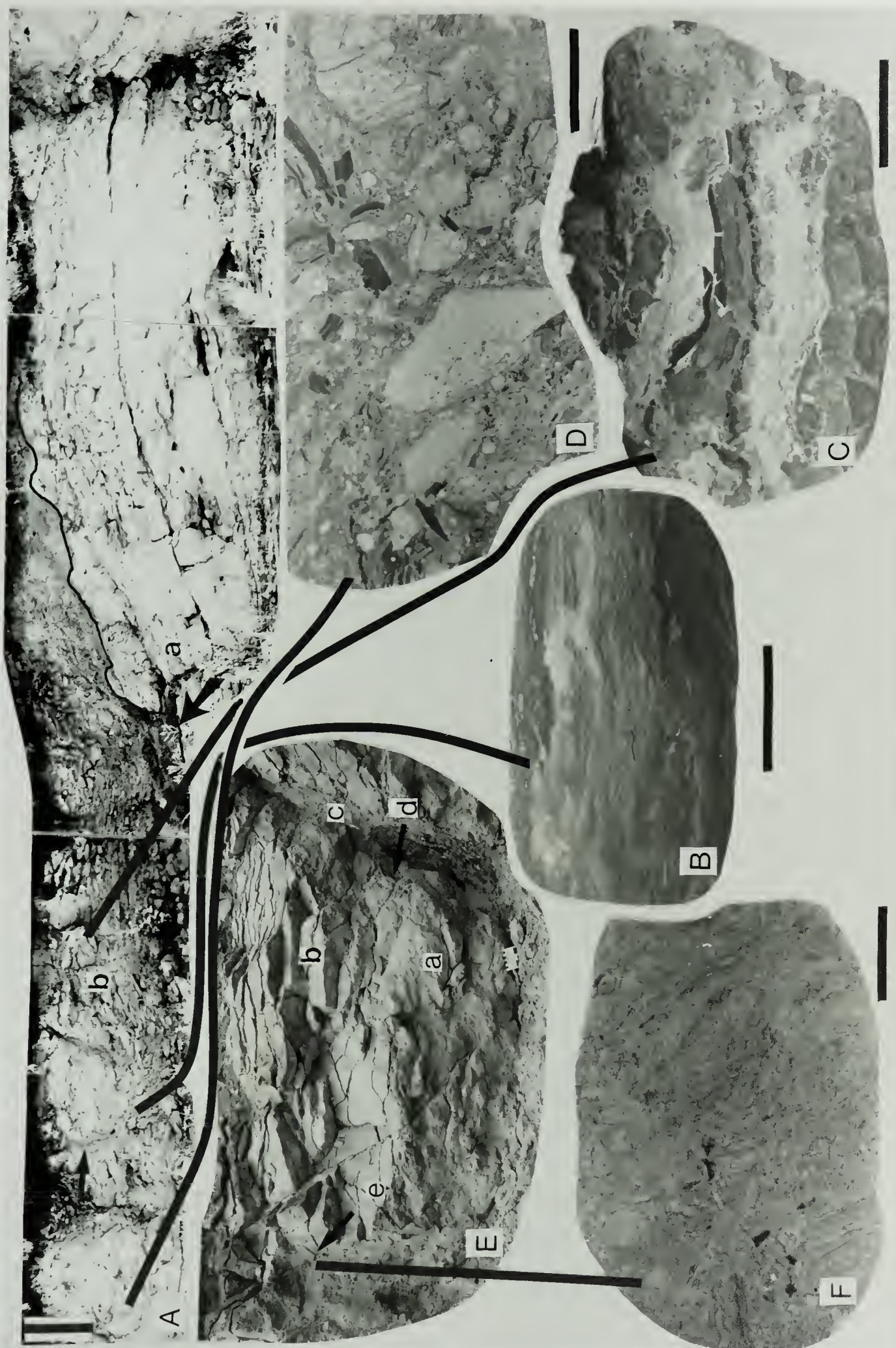
33B) Hand specimen of lithofacies 2c showing interlaminated texture of limestone (dark) and dolostone (light). Scale bar equals 2 cm

33C) Hand specimen of lithofacies 1d showing convoluted and distorted texture. Scale bar equals 2 cm.

33D) Hand specimen of lithofacies 8 showing how the Slave Point Formation becomes progressively more brecciated westward. The sample is a particulate rubble packbreccia. Scale bar equals 2 cm.

33E) Field photograph of a large block of lithofacies 2c (shown by arrow in A) from section PK. The lower and upper half of the block is limestone (a) and dolostone (b), respectively. Lithofacies 7 extends around the block in a pipe-like fashion (c) and also extends into the block (d). Vertical slickolites occur on the edge of the block (e). Scale card at base of block.

33F) Hand specimen of lithofacies 7 that was taken from pipe that extends around the block of lithofacies 2c. The breccia is comprised namely of dolostone. Scale bar equals 2 cm.



33E). This block is graded from limestone at the bottom to pure dolostone at the top. In the middle, along the outer edge of the block, well defined vertical slickolites are developed, suggesting that the block was moved vertically downward. A brown fine grained breccia with white clasts (lithofacies 7, Fig. 33F) completely encases the exposed portion of the block, and extends into the block, thereby breaking it apart. It appears that the block was isolated from the rest of the strata due to the formation of the lithofacies 7 pipe. Similar events produced another isolated block a few meters to the west. The strata below the more easterly isolated block are slightly more brecciated than in the surrounding area, thus suggesting that localized collapses occurred beneath these blocks.

Blocks of lithofacies 2c (mixing variety), are abundant towards the west beneath the base of the cliff that contains sections PK and PL (Figs. 24D, 24E, 24F, 24H). There, these blocks, along with massive dolostone and limestone blocks occur in lithofacies 8. Clasts of lithofacies 2c (mixing variety), that occur in lithofacies 8, were seen along the Peace River banks for most of the area between the Peace Point boat-launch-road to the eastern-most limit of the study area. These clasts display a gradational fabric between limestone and dolostone. Rims of either lithology form around the other with mixing of the two lithologies

decreasing toward the center of the clast. In places, limestone is rimming large bedded blocks of dolostone. Where the limestone is lithologically homogeneous, it tends to "break up" internally. That is, stylolites define individual clasts of limestone. The limestone at this stage is not brecciated, but the stylolites cause the limestone to look brecciated (Fig. 24A). Elsewhere, this process continued and the limestone did become brecciated, generally due to dolomite replacement along the stylolites. Elsewhere, small isolated pockets of lithofacies 7 formed where there was extensive mixing of dolostone and limestone (Figs. 24D, 24E). The process of dolostone and limestone mixing, seen to completion, forms lithofacies 7 as an end product. This process is similar to the alteration of chalk to a brown, rubbly, insoluble residue under a paleokarst surface near Oxford, South Island, New Zealand (Van Der Lingen et al., 1978). The mixing process seen in this study area is also thought to be a product formed beneath a subaerial exposure surface. Clasts of the mixing variety (lithofacies 2c) are found throughout many of the other lithofacies, however, these clasts are most common in lithofacies 8.

Moving away from the west end of section J, lithofacies 8 becomes the dominant lithology. This lithofacies also extends down below the base of the cliff. Westward, the degree of mixing and collapse

increases, thus showing the gradational transition of lithofacies 2 into lithofacies 8.

Lithofacies 8 changes laterally into lithofacies 9, where lithofacies 8 can be easily differentiated from the more recessive lithofacies 9. The contact between lithofacies 8 and lithofacies 9 is at the west end of section PK and in the breccia below the base of the cliff. The contact is gradational because mixing of the more shaly, greener colored, lithofacies 9 and lithofacies 8 does occur. In places throughout the study area, lithofacies 8 is greener, thereby suggesting that the overlying lithofacies 9 was not too far above the exposed level of outcrop, stratigraphically speaking.

SECTION PT

Section PT is almost identical to section PK in terms of the distribution of the lithofacies. On the very edge of section PT, the Slave Point Formation unconformably overlies bedded evaporites of the Fort Vermilion Formation (Fig. 34A). In section PT, the Slave Point Formation is highly contorted and slightly brecciated, but bedding can be followed laterally, parallel to the Fort Vermilion erosional surface. The Slave Point Formation becomes increasingly more brecciated towards the top of the unit with the uppermost stratiform breccia extending westward where it forms a thicker, non stratiform breccia. Again, a gradational brecciation process is evident from the

FIGURE 34

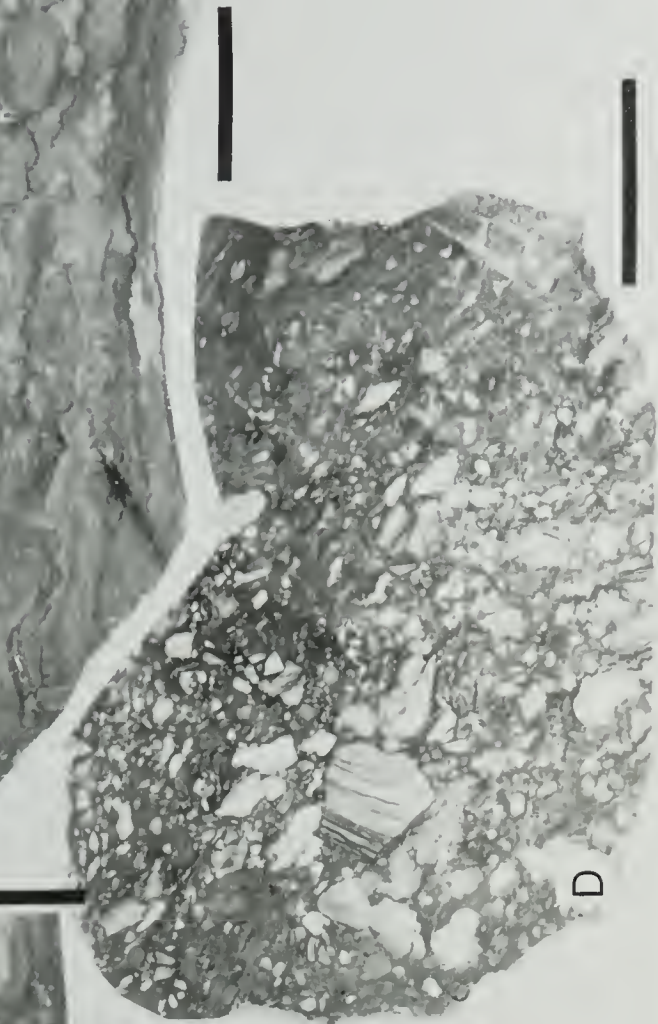
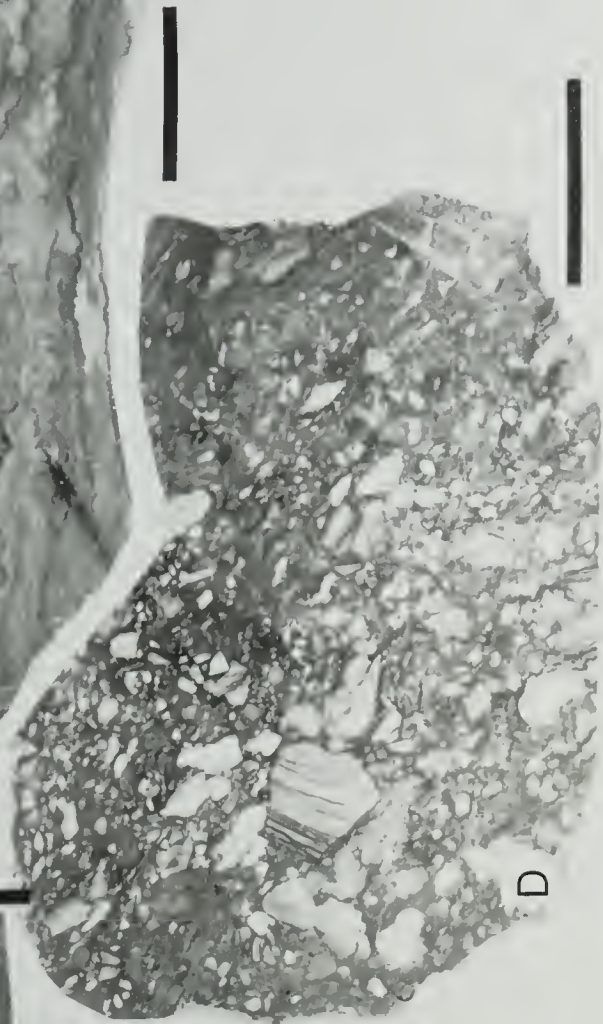
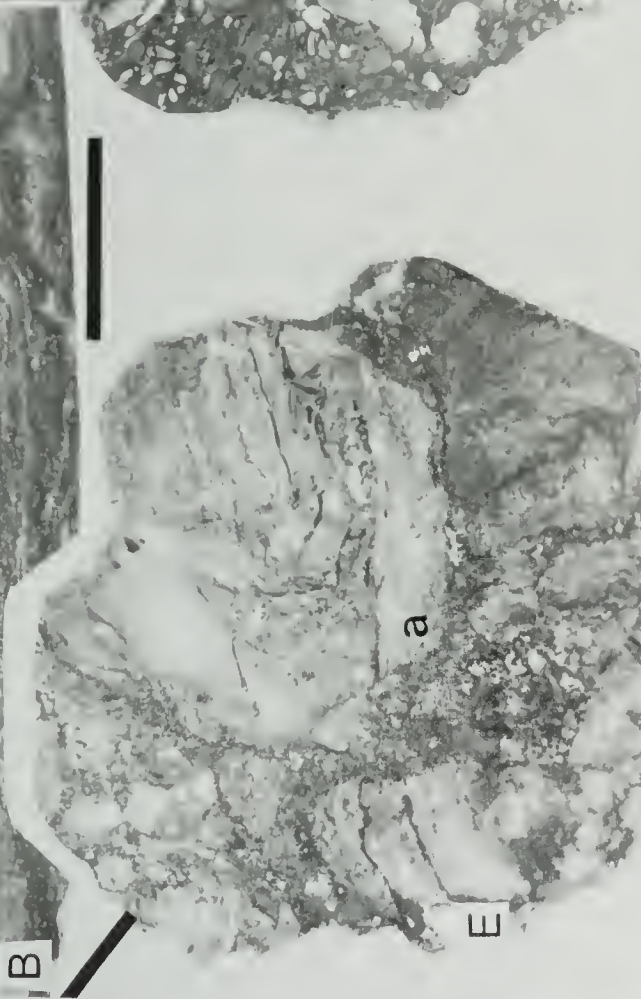
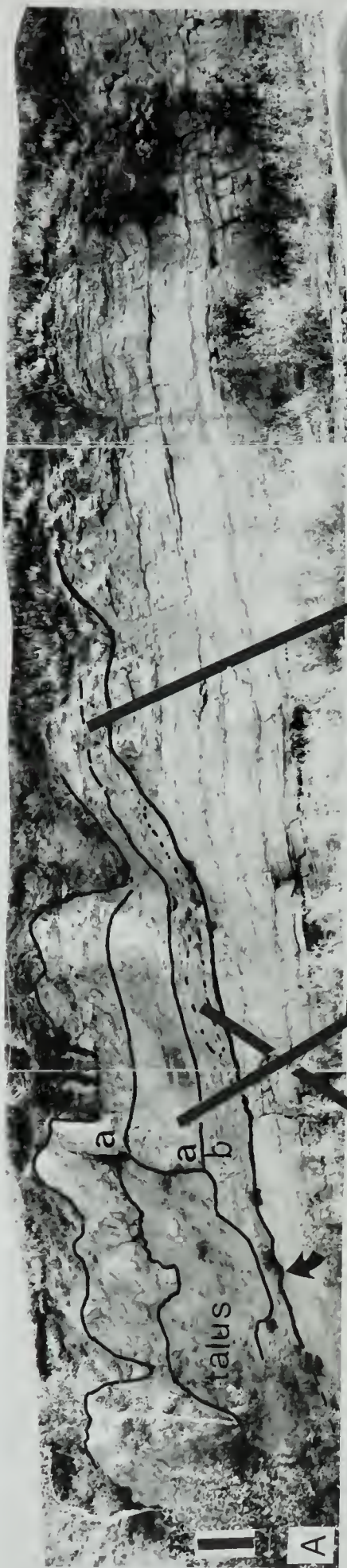
34A) Field photograph of section PT showing the Fort Vermilion Formation/Slave Point Formation angular unconformity (solid black line). At the lowest point of truncation of the Fort Vermilion strata (at black arrow) there are 17.3 m of Fort Vermilion strata missing. Note that brecciated strata (a) overlie nonbrecciated strata (b) with the Slave Point strata increasingly more brecciated westward (to the right of the photo). This area occurs in the very eastern edge of a major collapse zone with the center of the collapse to the left of the photo. Scale bar equals 1.5 m.

34B) Hand specimen of lithofacies 2a from the unit directly overlying the unconformity. Lithofacies 2a occurs as a relict lens in a unit comprised mainly of lithofacies 2b and 2c. Scale bar equals 2 cm.

34C) Hand specimen of lithofacies 2c/2d. Interlaminated dolostone (light areas) and limestone (dark areas) was beginning to develop a ropey texture. Scale bar equals 2 cm.

34D) Hand specimen of lithofacies 7 from second unit above the unconformity. The occurrence of white, laminated/ nonlaminated, chalky, dolomitic, limestone clasts in a dark brown argillaceous limestone matrix is typical of terra rossa deposits. Scale bar equals 2 cm.

34E) Hand specimen of lithofacies 8 taken from breccia west of photo, however, the sample is also representative of the uppermost unit in A. Note the sugary texture (a) of the dolostone clasts and the particulate rubble packbreccia lithology. Bar scale equals 2 cm.



outcrop.

Directly overlying the Fort Vermilion/Slave Point Formation unconformity is a 1.1 m thick unit consisting of lithofacies 2a, 2b, 2c and 8. This unit becomes rubbly and brecciated towards the top and west, locally displaying flaggy bedding. Lithofacies 2a occurs as a relict, unaltered lenticular slab near the base of the unit (Fig. 34B), with lithofacies 2c and 2d common towards the top and the east (Fig. 34C). Lithofacies 8 forms on the western edge of the unit, adjacent to the major collapse zone showing that the unit becomes laterally brecciated westward.

A 1.85 m thick interval of lithofacies 7, that has retained a bedded texture, overlies the lower unit with a gradational contact (Fig. 34D). Unlike section PK, where lithofacies 7 occurs in pipes and pockets, lithofacies 7, in section PT forms as part of a vertical brecciated succession with measurable dimensions.

Lithofacies 7 is overlain by a 1.8 m thick unit of lithofacies 8 (Fig. 34E). The contact between the two lithofacies is gradational. The breccia thickens towards the west, away from lithofacies 1, towards the center of a collapse zone comprised of lithofacies 8.

GRADATIONAL BETWEEN LITHOFACIES 6 AND 8

Lithofacies 6 and lithofacies 8 are gradational breccia lithofacies with gradational contacts developed between the two lithofacies throughout the study area (sections PA, PB,

PC, PD, PT and PW). The gradational aspect of the upper contact is seen in the decrease of gypsum clasts and in the increase of dolostone clasts from lithofacies 6 through lithofacies 8. The gradational aspect is also reflected in the similarity of the degree of brecciation and collapse between lithofacies 6 and 8, from one exposure to another. Where lithofacies 6 has undergone a greater degree of brecciation (collapse) so has lithofacies 8, with the degree of brecciation being proportional between the two lithofacies at any given outcrop. The degree of brecciation is also proportional to the degree of collapse, with the larger collapses reflected by a more chaotic texture in the breccia.

C. UNCONFORMITIES

FORT VERMILION/SLAVE POINT UNCONFORMITY

In the Fort Vermilion Formation (lithofacies 1), in section PU, a marker bed aided in the unit measurement. However, the presence of this marker bed also helped to determine the relationship of the Fort Vermilion/Slave Point contact. The maximum measured vertical section of Fort Vermilion strata in the study area is 23 m (section PU) with 16 m measured above the marker. Lithofacies 1 beds were traced downsection from the marker and westward towards the Fort Vermilion/Slave Point contact (located at the western end of section PU). There, a total of 7.2 m occurs below the marker. Individual beds downstream of the marker are

obviously truncated at the Fort Vermilion/Slave Point Formation contact (Fig. 34A). From the base of the Fort Vermilion Formation to the lowest point of truncation of the Fort Vermilion strata by the Slave Point Formation, there are 5.7 m of Fort Vermilion strata. Thus, there are 17.3 m of lithofacies 1 missing at the Fort Vermilion/Slave Point Formation contact (Fig. 34A). Following the abrupt and highly irregular contact east, the strata that are is not truncated become relatively younger. The Slave Point Formation is bedded parallel to the truncation surface.

In section PJ, 10.05 m of lithofacies 1 occurs, with the base of the section occurring directly below the Fort Vermilion/Slave Point contact. From the base to the Fort Vermilion/Slave Point contact a minimum of 3.5 m was measured. To the east (upsection), 6 m of lithofacies 1 occurs from the base to the 0.2 m thick marker bed, with 3.85 m occurring above the marker bed to the top of the exposed section. Thus, 6.55 m of lithofacies 1 is missing at the Fort Vermilion/Slave Point contact, located at the western end of section PJ (Fig. 33A). There, the contact is scalloped and highly irregular and the overlying Slave Point Formation is bedded parallel to the contact.

Angular contacts, where the Fort Vermilion Formation is distinctly truncated, occur in sections PJ, both the eastern and western ends of PM, and in domes of lithofacies 1, in section PX. In section PX, the angular aspect of the contact is extremely obvious because folded strata of lithofacies 1

are truncated by a horizontal erosion surface (Fig. 35A). In section PE there is truncation of lithofacies 1 on a smaller scale, with oxidized, brick red weathering, mudcracks occurring in lithofacies 2b, that directly overlies lithofacies 1 (Figs. 35B, 35C, 35D).

The top of lithofacies 1 is lithologically different from one exposure to another. In section PE, the unit is a gypsum that contains a nodular anhydrite. However, in section PH, a greenish brown calcareous mudstone occurs between lithofacies 1 and lithofacies 2 of the Slave Point Formation. However, on the east limb of the anticline, in section PH, the claystone changes laterally to gypsum.

INTERPRETATION

The missing section, the truncation of the Fort Vermilion Formation at the contact, the scalloped and irregular nature of the contact, the presence of dessication cracks directly above the contact, the presence of a claystone along the contact and a laterally varying lithology directly below the contact, clearly mark the angular unconformity between the two formations. The origin of this contact is due to either 1) an angular unconformity produced by uplift and subsequent erosion, or 2) an angular unconformity formed by interstratal dissolution of the gypsum. Several factors discount the latter possibility.

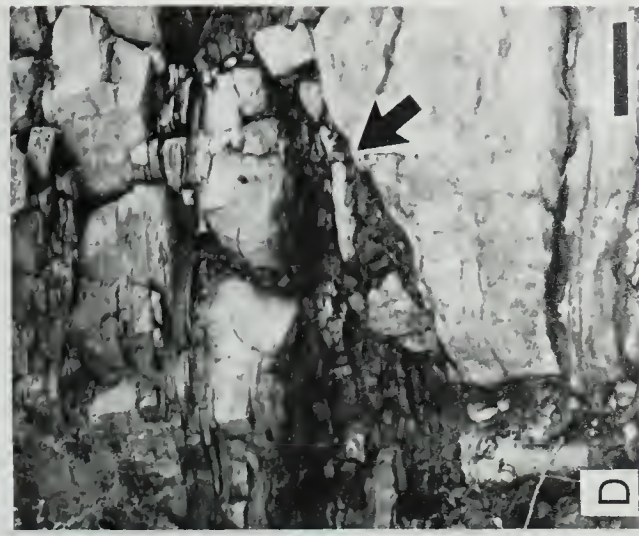
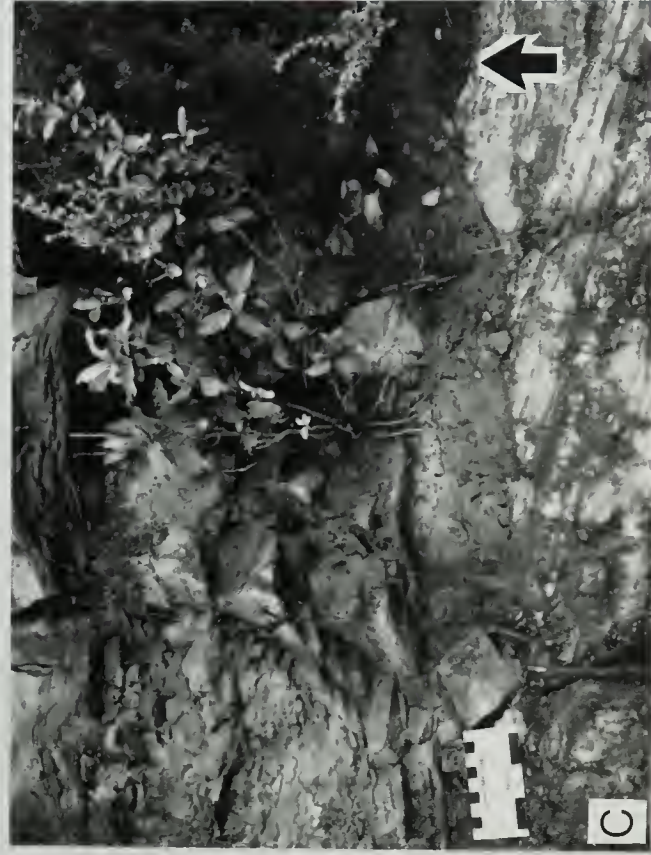
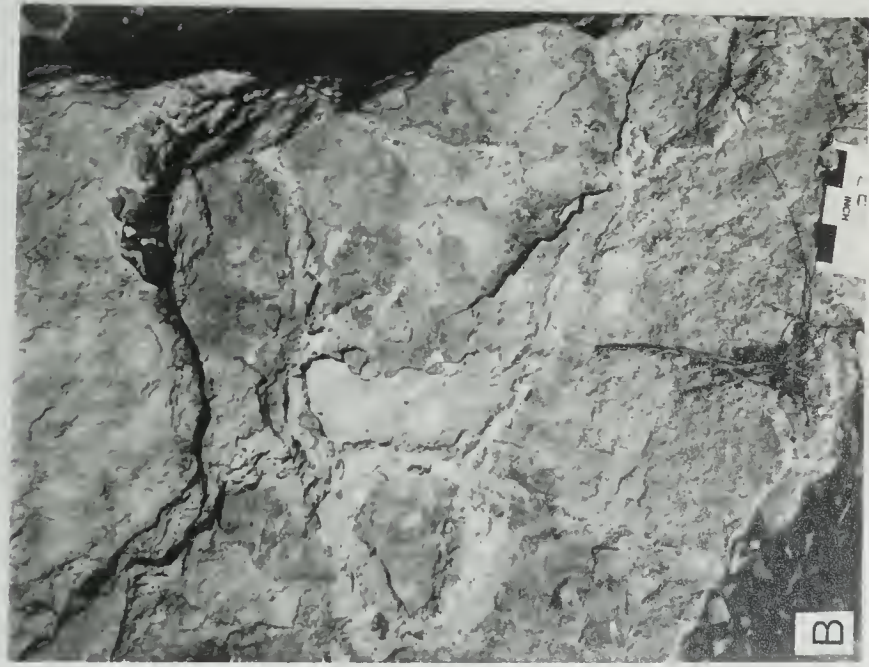
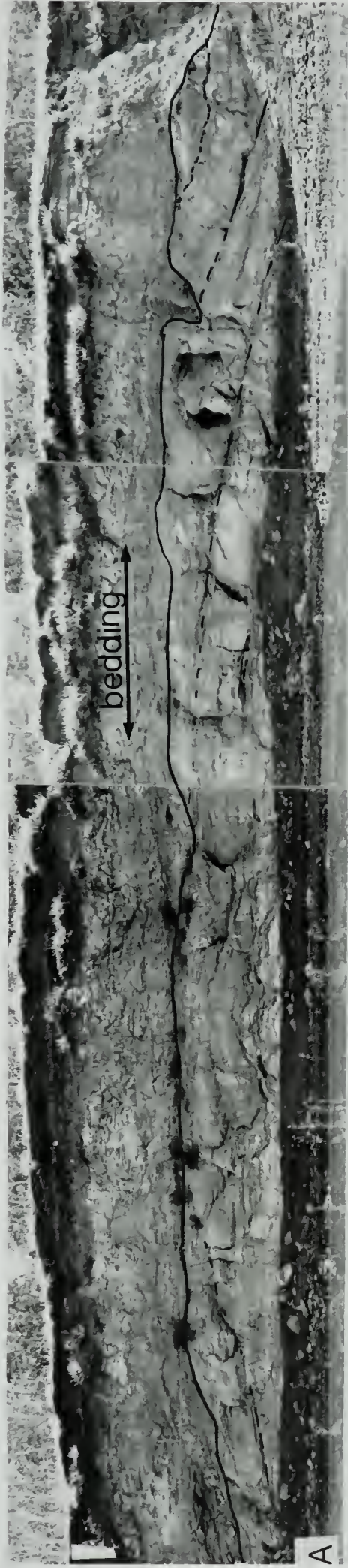
1. 17 m of lithofacies 1 have been removed and the overlying strata of the Slave Point Formation are

FIGURE 35

35A) Field photograph of the lithofacies 1 (Fort Vermilion Formation)/lithofacies 2b/2c (Slave Point Formation) contact (solid dark line). Note that bedding (dashed lines) in the evaporite is truncated at the contact with the overlying Slave Point Formation bedded parallel to the contact. The above evidence suggests that lithofacies 1 was domed prior to an angular unconformity that developed above the evaporite. Scale bar equals 1.5 m.

35B) Field photograph of highly oxidized, brick red mudcracks, taken directly above the Fort Vermilion Formation/Slave Point Formation unconformity (section PE).

35C & D) Field photographs of the Fort Vermilion Formation/Slave Point Formation angular unconformity (section PE) showing the shaly, highly oxidized mudcrack-containing-limestone directly overlying. Note how lithofacies 1 is truncated at the erosion surface (arrow). Scale bar in D equals 10 cm.



bedded. If the truncation surface was due to interstratal dissolution then the overlying rock would have collapsed down, thus having caused "complete" brecciation of the overlying Slave Point strata. The Slave Point strata, however, in many areas is not brecciated or not severely brecciated above lithofacies 1, suggesting that collapse did not occur. Collapse did occur at a later time (collapse 2) which resulted in the slightly brecciated texture seen in some exposures.

2. Generally, dissolution of a unit is along bedding planes. Any break produced by interstratal dissolution would probably be represented in the form of a disconformity. In this case, the truncating surface cuts across bedding thus developing an angular discordance. The Slave Point strata are bedded parallel to the discordance.
3. If there was interstratal dissolution of lithofacies 1 there should be a gypsum breccia developed between the top of the bedded Fort Vermilion strata and below the Slave Point Formation. However, there is no gypsum breccia between these two lithologies. Dissolution of lithofacies 1 has produced gypsum breccia elsewhere in the study area, but this breccia formed after deposition of the shale of the Peace Point Member and is therefore not related to the Fort Vermilion/Slave Point contact.

4. Interstratal dissolution of a unit can occur in any lithology, which can be representative of any environment. Therefore, the presence of dessication cracks neither proves nor disproves interstratal dissolution. However, the presence of dessication cracks in lithofacies 2b, that directly overlies the unconformity, represents very shallow water to supratidal conditions. Such an environment would be expected if the Devonian Sea had transgressed back onto an area which was uplifted and eroded.

Based on the above evidence the boundary between the Fort Vermilion Formation and Slave Point Formation, in the Gypsum Cliffs area, is an angular unconformity, produced by uplift and subsequent erosion of the Fort Vermilion strata at the end of Fort Vermilion deposition.

THE SLAVE POINT FORMATION/PEACE POINT MEMBER UNCONFORMITY

The upper contact of the Slave Point Formation was not seen in the study area but the following observations are suggestive of subaerial exposure at the end of Slave Point deposition. Most of the lithofacies comprising the Slave Point strata are representative of diagenetic products formed at, or just below, the subaerially exposed surface of the Slave Point Formation.

A subaerial exposure surface is suggested by the chemical instability of rocks of the Slave Point Formation. Extensive dissolution and/or recrystallization of the Slave

Point strata produced lithofacies 2b, that was later lined with shale of the Peace Point Member and Cretaceous sandstone. The lining of these vugs with shale clearly shows that the diagenesis had to occur prior to shale deposition. Diagenesis at that time also involved severe mixing and chemical replacement of the Slave Point Formation, thus forming lithofacies 2c.

The most obvious evidence of subaerial exposure is in the fact that the Slave Point Formation contains strata that are similar to terra rossa (Lithofacies 7) and calcrete (portions of Lithofacies 8). To what extent lithofacies 8 is calcrete is not known, for clasts of calcrete were incorporated into lithofacies 8 along with other lithofacies, during collapse 2. Extensive diagenesis of lithofacies 2 formed at approximately the same time as the calcrete. The calcrete portion of lithofacies 8, lithofacies 7 and lithofacies 2 may be representative of a vertical soil profile that had formed from alteration of the Slave Point Formation.

Throughout lithofacies 8, clasts in this breccia contain additional evidence of very shallow water to subaerial conditions. Such clasts contain intraformational breccia and mudcracks (Fig. 28F).

Kahle (1978) and Esteban and Klappa (1983) listed numerous criteria for evidence of subaerial exposure. Of those listed, the following criteria occur in the strata exposed at Peace Point:

1. Blackened lithoclasts
2. Breccias
3. Caliche
4. Chalky limestone
5. Clays localized along discontinuities such as bedding planes, fractures or both.
6. Clotted or peloidal micrite, with microspar channels and cracks
7. Coated grains
8. Flowstone and coarse druse covering irregular surfaces - occurs rarely on clasts in lithofacies 8
9. Evidence for in-place alteration and replacement of the original rock or sediment.
10. Evidence for leaching such as caverns, vugs, leached ooids, or fossils.
11. Fissures (filled with sandstone)
12. Fractures on a large or small scale.
13. Joints that display evidence of solution-widening.
14. Karstic collapsed structures in the form of rubble filled sinkholes, or rubble filled fissures or caverns that may be roughly spherical to highly irregular in shape.
15. Macroborings
16. Mudcracks
17. Oxidized and blackened zones
18. Staining of sediment or rock in shades of red, orange, and black

19. Channels filled with sandstone
20. Clay cutans
21. Alveolar texture
22. Rillenkarren - possibly ancient in the study area, but may also be due to Recent weathering.
23. Floating texture
24. Tepees
25. Shallowing upward sequence culminating in nonspecified exposure surface
26. In place, non tectonic, fracturing and brecciation.

Each of the above observations, along with observations by Norris (1963) and Norris and Uyeno (1983), does not in itself confirm the presence of subaerial exposure; however, together there is sufficient evidence to suggest that subaerial exposure occurred at the end of deposition of the Slave Point Formation, prior to deposition of shale of the Peace Point Member.

THE PRE-CRETACEOUS UNCONFORMITY

A major unconformity exists between the Upper Devonian (Frasnian Age) strata and the Cretaceous (Albian Age) strata in northeastern Alberta (Canadian Society of Petroleum Geologists Highway Map of Alberta). On a southwest to northeast cross-section across Alberta, the Upper Devonian Formations (Ireton Formation and Grosmount Formation) are truncated towards the northeastern corner of Alberta.

Deposition of the Ireton Formation and the Grosmount Formation or any other post-Peace Point

Member/pre-Cretaceous sediment did not occur in the study area because there are no lithologies indicative of post-Peace Point Shale or Pre-Cretaceous strata in the breccias or in the vuggy limestone of lithofacies 2b. Lithofacies 1 and 6 are not observed in lithofacies 10 because they are relatively older than the sandstone collapsing event and they are not involved with the brecciating event producing lithofacies 10.

One lithology of Cretaceous age does occur in lithofacies 2. Lenticular beds of the calcareous medium grained sublitharenite represent the first strata deposited in the area since deposition of the shales of the Peace Point Member. This sandstone is identical to sandstone sampled from the Cretaceous outlier to the north (Fig. 32A), with respect to lithology and flora contained and has been equated to the McMurray Formation (Singh, Appendix 5). The occurrence of this sandstone in lithofacies 2b, along with no evidence of deposition between the sandstone and the Peace Point Member suggests that a major unconformity occurs between deposition of the two lithologies. However, based on the fact that the Peace Point Shale was washed into vugs in lithofacies 2b, the unconformity probably began with subaerial exposure developed at the end of the Slave Point Formation, prior to shale deposition of the Peace Point Member. Deposition of this shale reflects a minor transgression of the sea during the Devonian, that is relatively insignificant when compared to the geological

time represented by the interval of subaerial exposure. This interval of subaerial exposure lasted into the early Cretaceous when the calcareous sandstone of Upper Barremian age was washed into lithofacies 2. Further deposition of Cretaceous strata continued up into Maestrichtian time when collapse 3 occurred.

D. RELATIVE AGES OF THE LITHOFACIES

Lithofacies 1, 2, 3, 7 and the calcrete portion of lithofacies 8, mark the initial depositional events and subsequent diagenetic changes that were produced after deposition. Diagenetic alteration was extensive and occurred under a paleo-subaerial exposure surface that was developed prior to shale deposition of the Peace Point Member.

The oldest lithofacies in the study area is lithofacies 1. An angular unconformity was produced on the top of lithofacies 1 after it was uplifted and folded. Doming was prior to deposition of the Slave Point Formation because the Slave Point strata are bedded parallel to the unconformity which cuts horizontally across previously folded strata of lithofacies 1 (Fig. 35A).

Lithofacies 2 was deposited next with lithofacies 2a forming the oldest lithofacies in lithofacies 2. Lithofacies 2b and 2c are younger deposits for they represent later diagenetic alterations of lithofacies 2b. Lithofacies 2d is related to stress developed on the edges of the breccias, near the outer areas of a collapsed zone and had formed

during collapse 2.

Lithofacies 7 and the calcrete portion of lithofacies 8 may be the next youngest lithofacies formed. However, the relative age of these lithofacies when compared to the age of lithofacies 2b and 2c is speculative. These lithofacies are probably synchronous for they are all products produced by alteration of the original Slave Point strata and together possibly represent a vertical karst/calcrete profile through the Slave Point Formation. However, the calcrete is probably slightly younger for it caps the karst profile and probably represents formation under slightly different climatic conditions.

Lithofacies 4 formed just prior to lithofacies 2b, 2c, 7 and the calcrete portion of lithofacies 8 because the majority of the clasts in lithofacies 4 represent the original lithology of the Slave Point Formation and have not been as diagenetically altered. Minor amounts of lithofacies 2b and 2c do occur in lithofacies 4, but were probably incorporated into the edges of the breccia during collapse 2, when lithofacies 4 went through a second collapsing phase. Lithofacies 4 occurred prior to shale deposition of the Peace Point Member because there is no shale observed in this lithofacies.

Collapse 1 is the event responsible for the formation of lithofacies 4, which is representative of paleo-sinkholes formed prior to deposition of the Peace Point Shales and collapse 2. To the west of the study area, Norris (1963)

suggested that the collapse of the Slave Point Formation occurred prior to shale deposition of the Peace Point Member, with the shales later deposited into these depressions. There, the shale was obviously deposited into a sinkhole that had previously formed. However, in the present study area such relationships are not apparent because the shale was brecciated later during collapse 2, thereby forming the shaly breccia of lithofacies 5.

Lithofacies 4 contains clasts that are themselves brecciated, indicating that at least two phases of brecciation occurred. The first phase of brecciation was associated with minor development of sinkholes at the end of Slave Point deposition (collapse 1). The second phase of brecciation occurred later during collapse 2.

Physically, Lithofacies 4 appears to cut vertically through lithofacies 7 and 8. But, for reasons already given, the pipes are older than these lithofacies, therefore, are not cutting lithofacies 7 and 8. One explanation is that lithofacies 4 remained as vertical pillars, while the other lithofacies collapsed down around the pipes. Or, the pipes were inert objects that did not mix that much with lithofacies 7 and 8 when they were incorporated into lithofacies 8 during collapse 2.

Shale of the Peace Point Member is considered younger than the dissolved, diagenetically altered, strata of the Slave Point Formation since the shale lines vugs and bedding surfaces in lithofacies 2b. The diagenetically produced

texture of lithofacies 2b would have to precede the deposition of the shale.

The amount of shale of the Peace Point Member in each breccia, that is formed as a result of collapse 2, is gradational with respect to the relative age of the breccia. This statement excludes the shale that occurs in lithofacies 5 (the shaly breccia) for it forms as concentrated pockets irrespective of the relative ages of the breccias. Lithofacies 6, 8, and 9, the next youngest units, represent semi-layered breccia deposits that formed during collapse 2. Shale of the Peace Point Member was originally deposited above the stratigraphic level of these breccias, but upon collapse 2 the shale was incorporated into all three breccias. The shale content decreases down through lithofacies 9 to 6 as the stratigraphic distance from the original stratigraphic position of the Peace Point Member is increased. As is expected, lithofacies 6 contains the least amount of shale and lithofacies 9 the most.

Collapse 2 involved the collapse of many lithofacies. Some lithofacies were changed whereas others were not. For example lithofacies 1 and the Peace Point Shale collapsed thus forming a major part of lithofacies 6 and lithofacies 5, respectively. Lithofacies 2, 3, 4, 5, 7 and the calcrete portion of lithofacies 8 remained distinct lithologies, however during collapse they were incorporated into lithofacies 8. Lithofacies 9 was a newly produced lithofacies into which a good percentage of the shale of the

Peace Point Member was incorporated.

Regardless of how or why the lithofacies diagenetically formed, the fact remains that they all collapsed together during collapse 2. The gradational aspect of the lithofacies produced during collapse 2, with respect to lithofacies contacts, textures and lithologies, supports the theory that the brecciation (collapse 2) occurred as a major single event that accounts for most of the breccias observed at Peace Point.

Collapse 3 formed the youngest breccia in the study area (lithofacies 10). This breccia was formed syn-Late Cretaceous and prior to glaciation.

VIII. MODEL OF BRECCIATION

A. ENVIRONMENTAL CONTROLS AND BRECCIA TYPES

KARST/CALCRETE PROFILE

In the Gypsum Cliffs area there is evidence to suggest that a karst profile formed during the Devonian. There is also evidence to suggest that a calcrete capped the Slave Point strata above the Karst profile. Karst and calcrete are generally considered two end members of a gradational spectrum where the formation of the end members is governed by factors of climate, soil cover, substrata and time (Read and Grover, 1977; Harrison, 1977, Harrison and Steinen 1978).

The karst facies involves a net loss of calcium carbonate whereas, in the calcrete facies there is a zero balance between calcium carbonate dissolved and calcium carbonate precipitated or a net gain of calcium carbonate from an external source (Esteban and Klappa, 1983). "Karst is a diagenetic facies, an overprint in subaerially exposed carbonated bodies, produced and controlled by dissolution and migration of calcium carbonate in meteoric waters, occurring in a wide variety of climatic and tectonic settings, and generating a recognizable landscape" (Esteban and Klappa, 1983, p.22). Esteban and Klappa (1983) stated that the soil cover on karst profiles is varied, with terra rossa, laterites, or even calcrete forming. They stated calcrete can form during mature and late stages of karst

evolution. Also, calcrete is developed on karst in the Carboniferous limestone of the Derbyshire Block, England (Walkden, 1974) and on karst in the Gower Peninsula, South Wales (Wright, 1982). The presence of calcrete on karst is not common, but it can occur.

Sea level oscillations can complicate the karst profile by repeatedly changing the positions of the vadose and phreatic environments. Such changes can cause diagenetic histories which are virtually impossible to decipher. In contrast, sea level changes did not effect the Gypsum Cliffs area extensively for the area was high and dry until the Cretaceous. However, a minor transgression of the Devonian seas occurred when the shales of the Peace Point Member were deposited. This event, in terms of geological time, is insignificant when compared to the time interval represented by subaerial exposure. Therefore, the Gypsum Cliffs strata underwent a long, uncomplicated stretch of subaerial exposure after deposition of the Slave Point Formation.

During subaerial exposure a karst profile developed in both the evaporite lithology of the Fort Vermilion Formation and the carbonate lithology of the Slave Point Formation. With regression of the sea, the exposure surface and associated underlying vadose and phreatic environments would have moved vertically down through the strata. This karst profile matured until calcrete eventually developed on top of the profile. Because of the long period of subaerial exposure the development of the caliche could be due to

numerous factors, but the main factor was probably a change in climate. Associated features such as dissolution and cave production resulted in the three collapses and subsequent breccias and most of the lithofacies formed in the study area.

RELATIONSHIP OF THE LITHOFACIES TO A KARST/CALCRETE PROFILE

The Peace Point strata have obviously been altered by karsting. However, to take a generalized karst profile that has been divided into an "X" number of zones and superimpose it onto a given vertical sequence of rocks is difficult to do. The interaction of chemical and physical parameters that formed the karst profile is complex and is a study in itself. But, general comparisons of the lithofacies with a generalized karst profile (Esteban and Klappa, 1983, p. 4) will hopefully accentuate the premise that a karst profile does exist in the Gypsum Cliffs area.

Remnant lithologies, representative of the original lithologies of the Slave Point Formation, are lithofacies 2a and 3. Lithofacies 2b, 2c and 7 are the karst related lithofacies. Lithofacies 2c forms massive dolostones that generally overlies lithofacies 1, whereas lithofacies 2b forms a vuggy lithofacies, generally overlying lithofacies 2c. Lithofacies 2b is a result of extensive dissolution, whereas lithofacies 2c represents a net gain of material with dolostone replacing limestone. The sharp and planar contact developed between lithofacies 2b and 2c (section PE, Fig. 21A) may represent a break in the paleoenvironment

where vadose conditions and phreatic (or lower portions of the vadose zone) occurred above and below the contact, respectively. However, elsewhere, a gradational contact had formed where patches of lithofacies 2b and 2c occur in lithofacies 2c and 2b, respectively, thereby suggesting that the paleoenvironment was gradational.

The dissolved recrystallized texture of lithofacies 2b is typical of textures that occur in the vadose zone beneath subaerial exposure surfaces. For example, solution vugs in the Bertie Dolomite represent extensive fresh water dissolution in the vadose zone beneath the Silurian-Devonian disconformity (Kobluk et al., 1977). Kobluk et al. (1977) also added that vadose silt and sand were washed into some of the vugs. That situation is extremely similar to the sandstone that filled in lenses in lithofacies 2b.

Klappa and Esteban (1983) divided the vadose environment into an upper dissolution zone of infiltration and a lower percolation zone of downward water movement. In the infiltration zone, high levels of carbon dioxide cause dissolution, whereas, lower carbon dioxide levels can cause precipitation. The percolating downward water is in calcium carbonate-carbonic acid equilibrium by the time the percolation zone is reached, therefore having minor dissolution affects (Esteban and Klappa, 1983). The extensive vugginess of lithofacies 2b suggests that the carbon dioxide content was high. In the upper part of the percolation zone (lower vadose zone) localized areas of

vadose flow can form vugs and caves. However, in the lower 2 m of the percolation zone a capillary fringe directly above the water table allows for intense carbonate precipitation (Esteban and Klappa, 1983). Diagenesis in a capillary fringe environment may be responsible for the massive bedded deposits of dolostone of lithofacies 2c. The increasing concentration of this dolostone down to lithofacies 1 suggests that magnesium derived from the evaporite aided in forming the dolostone in the capillary fringe environment. The fact that the most concordant, thickest deposits of dolostone, directly overlie broad flat plateaus of lithofacies 1 (section PE) suggests that downward percolating meteoric water simply "ponded" on the impermeable lithofacies 1. The capillary fringe would be horizontal over a horizontally bedded evaporite and the sharp contact between lithofacies 2c and 2b, in section E, reflects this. Whereas, on the smaller anticlines of lithofacies 1 water flowed from the crests into lows between the anticlines thus producing only patchy discordant dolostone on the evaporite highs. The abundance of meteoric water in the lows caused extensive dissolution of lithofacies 1, which eventually led to the formation of caves. These caves are not seen in outcrop because they have since collapsed, thus forming lithofacies 6, 7, 8 and 9. Most cavern porosity and intense formation of subhorizontal caves form in the lenticular (upper phreatic zone) beneath the percolation zones (Esteban and Kappa, 1983). Caves in

lithofacies 1 may have formed in a similar zone.

Lithofacies 7 is a terra rossa that developed as a residual alteration product from the Slave Point Formation. This lithofacies marks the top of the karst profile; but, karsting did not stop with development of the terra rossa, nor with development of the calcrete directly overlying the terra rossa. Evidence of this, is where the karst and calcrete lithologies are in the resultant breccia, thus suggesting they formed prior to the brecciation event (collapse 2). Instead, dissolution in the lenticular zone of the phreatic environment was continually enlarging caves beneath the lows in lithofacies 1, which eventually led to collapse 2.

BRECCIA TYPES

The terminology related to breccias with a collapse origin is confusing. Terms such as karst breccia, solution breccia, collapse breccia and evaporite-solution-collapse breccia have been used. Various authors have described the above breccias, however there are no clear cut criteria for distinguishing the breccia types. Detailed descriptions of the breccias (where lithology, clast size parameters, clast shape, matrix to clast ratios are given) is common practice, but, really of little value when comparing one type of "collapse" breccia to another. Taking two breccias with similar collapse origins, the variations in the amount of vertical displacement and the original lithology of the strata are alone enough to render the breccias significantly

different. There is however, some textural and spatial evidence that is semi-typical of some of the breccias. However, distinction between breccia types, with respect to genesis, is generally related to the definition of the breccia type.

Solution breccia "is a mass of rock composed of angular to rounded fragments of rock that have accumulated by solution of surrounding or underlying carbonate" (Monroe, 1970, p. K16). Collapse breccia is a "Mass of rock composed of angular to rounded fragments of limestone or dolomite that has formed as a result of the collapse of the roof of a cave, or of an overhanging ledge" (Monroe, 1970, p. K6). In these two definitions the final product is basically the same. The major distinction between solution breccia and collapse breccia, as used in this study, is the genetic mode of formation. Breccias result from dissolution in the former and cave formation and subsequent collapse in the latter. In this study, both these terms will only refer to dissolution of a carbonate lithology. Some of the more distinctive textures are a coarsening upward of clasts in solution breccias (Robert, 1966) and a less chaotic texture, with less rotation of blocks, towards the top of collapsed breccias (Campbell, 1977).

In evaporite-solution-collapse breccias the mode of formation is identical to collapse breccias, except that dissolution of evaporite, not carbonate, has caused cave formation and subsequent collapse of the overlying strata.

The basic criteria for recognition of evaporite-solution-collapse breccia is the stratigraphic position of the breccias relative to significant evaporite deposits (Blount and Moore, 1969). Johnston (1974) formed a table listing diagnostic characteristics of evaporite-dissolution-collapse breccias and carbonate-dissolution-collapse breccia of the Bass Islands Breccia, southeastern Michigan. The table shows that there is really no substantial difference between the breccias with the exception of extent. The evaporite-dissolution-collapse breccias occur as sheets and pipes, whereas the carbonate-dissolution-collapse breccias form laterally less extensive lenses and pipes.

Karst breccias and solution breccias are similar, but differ with respect to clay lithologies, a fewer chaotically distributed fragments in karst breccia and the presence of an upward coarsening of clasts in solution breccias (Roberts, 1966). The karst deposits are discontinuous, whereas the solution-breccia beds can be traced laterally (Roberts, 1966). Kahle (1978) interpreted 50% of the Silurian breccias in the Murnee stone quarry to be paleokarst breccias developed by solution brecciation and solution collapse. This usage implies that solution collapse breccias are a form of karst breccia. Most breccias of dissolution and/or collapse origin were formed under a subaerial exposure surface as part of a karst profile and therefore all represent karst breccias.

In the Gypsum Cliffs area three collapsing episodes have produced breccias for slightly different genetic reasons. Those breccias not a result of collapse have already been described (lithofacies 7 and calcrete of lithofacies 8). The remaining breccias are a result of dissolution and subsequent collapse in a paleokarst profile, therefore are all karst related breccias. However, following Kahle (1978) the breccias will be further divided into genetic types.

COLLAPSE 1 - Dissolution of carbonate of the Slave Point Formation caused collapse 1. Whether the breccia (lithofacies 4) was associated with the formation of caves and subsequent collapse of the caves or simply due to brecciation by solution, is not certain. But, the fact that Norris (Norris and Uyeno, 1983) reported stratiform breccia beds at the base of bedded sequences of Slave Point strata suggests that lithofacies 4 is the result of dissolution, not collapse. Depending on a dissolution or collapse mode, the associated landforms were either paleo-sinkholes or paleo-dolines, respectively. Therefore, the breccia of lithofacies 4 can be named either a solution breccia or a collapse breccia, but is probably best named a solution breccia.

COLLAPSE 2 - Dissolution beneath lows of lithofacies 1 caused collapse of the overlying lithofacies. There is no evidence of caves in lithofacies 1 because they were destroyed upon collapse. However, associated textures, the

History

- 1 Lithofacies 1 deposited
- 2 Uplift/doming
- 3 Angular unconformity produced

Symbols

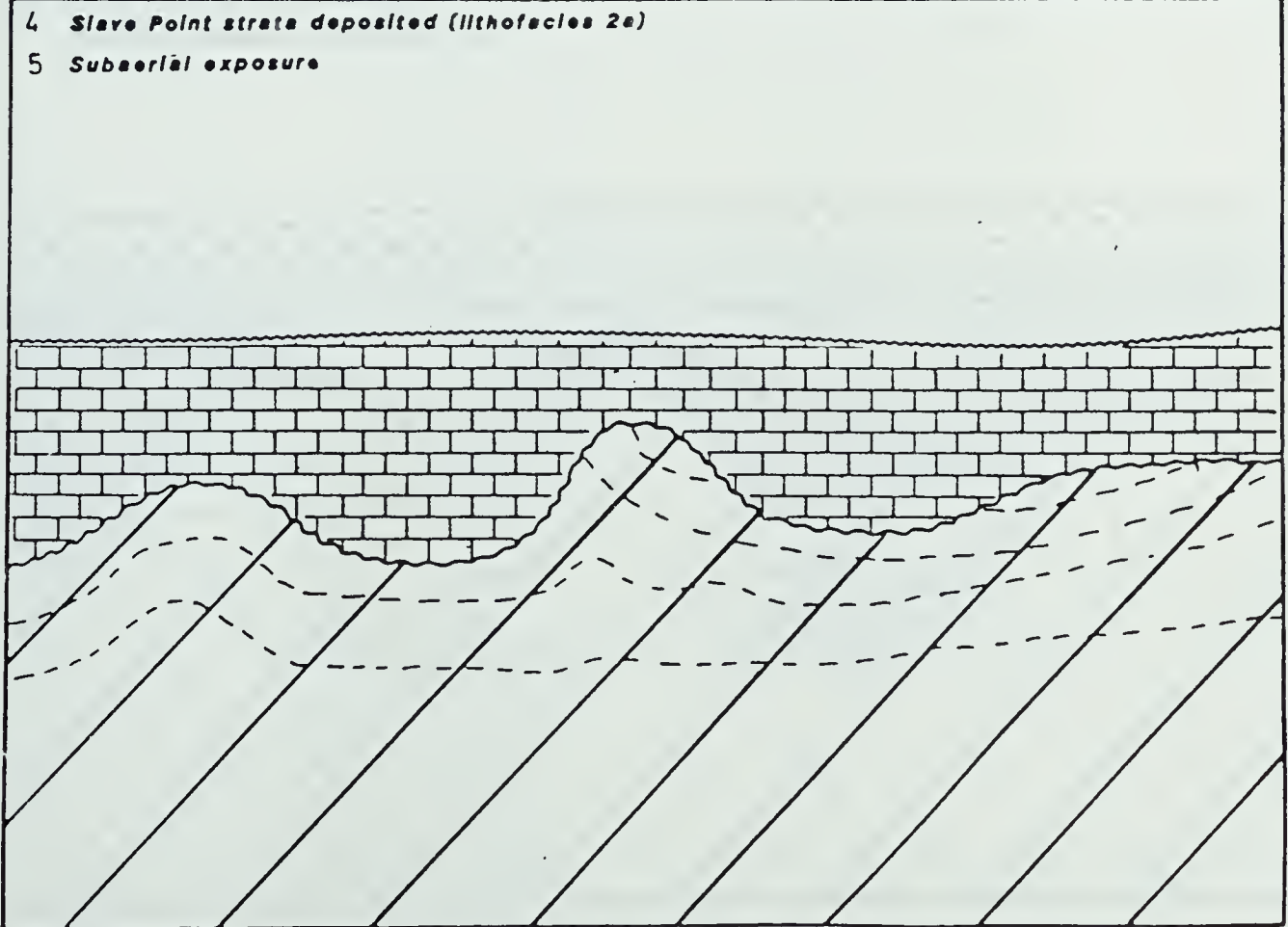
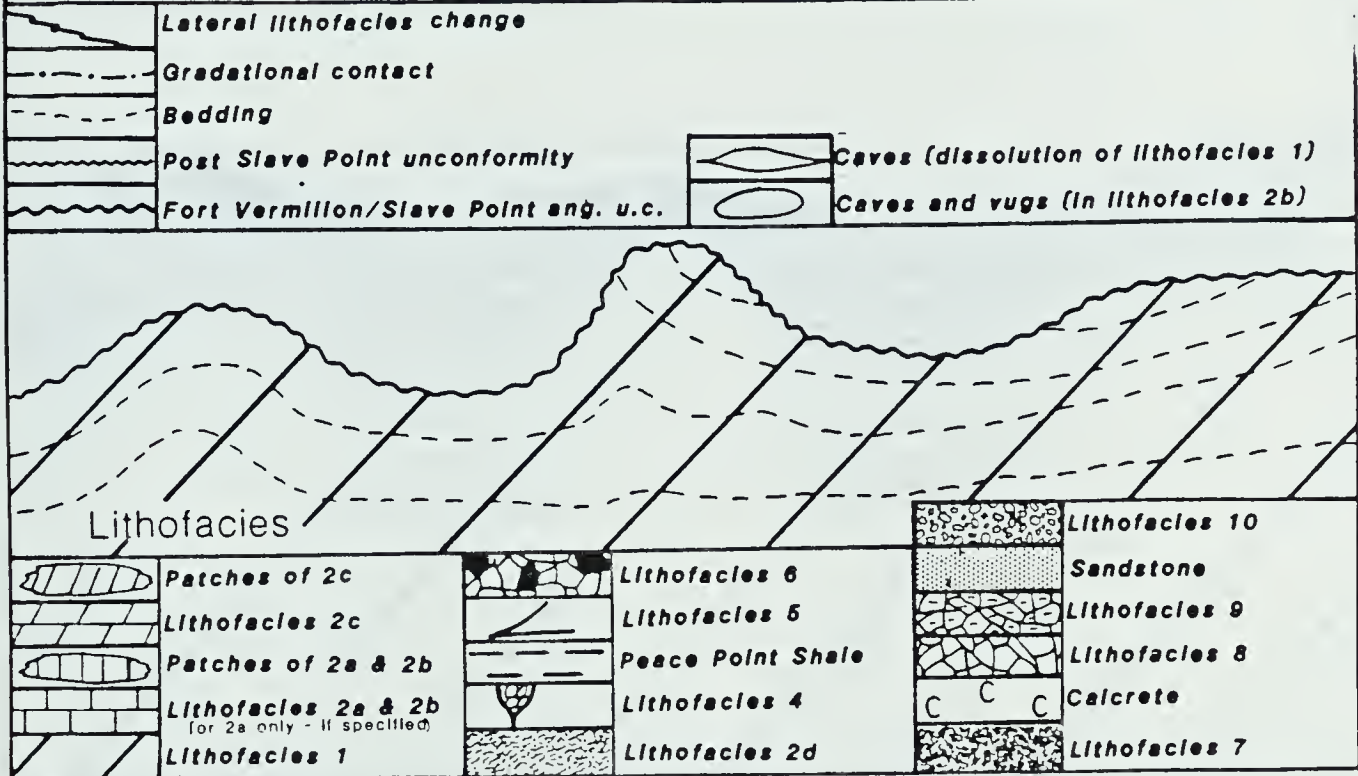


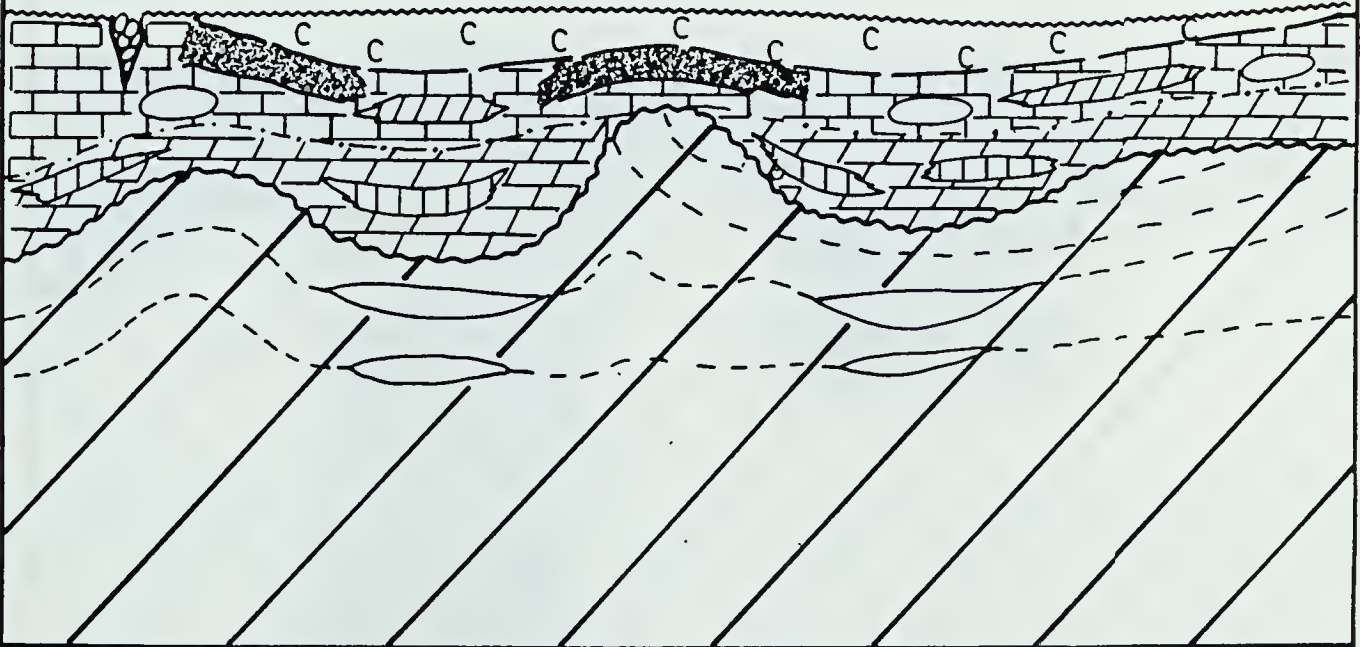
FIGURE 36. Schematic Drawings Depicting the Depositional and Diagenetic History of the Peace Point Area.

6 Collapse 1 - lithofacies 4 formed

9 Dissolution of lithofacies 1 (in lows)

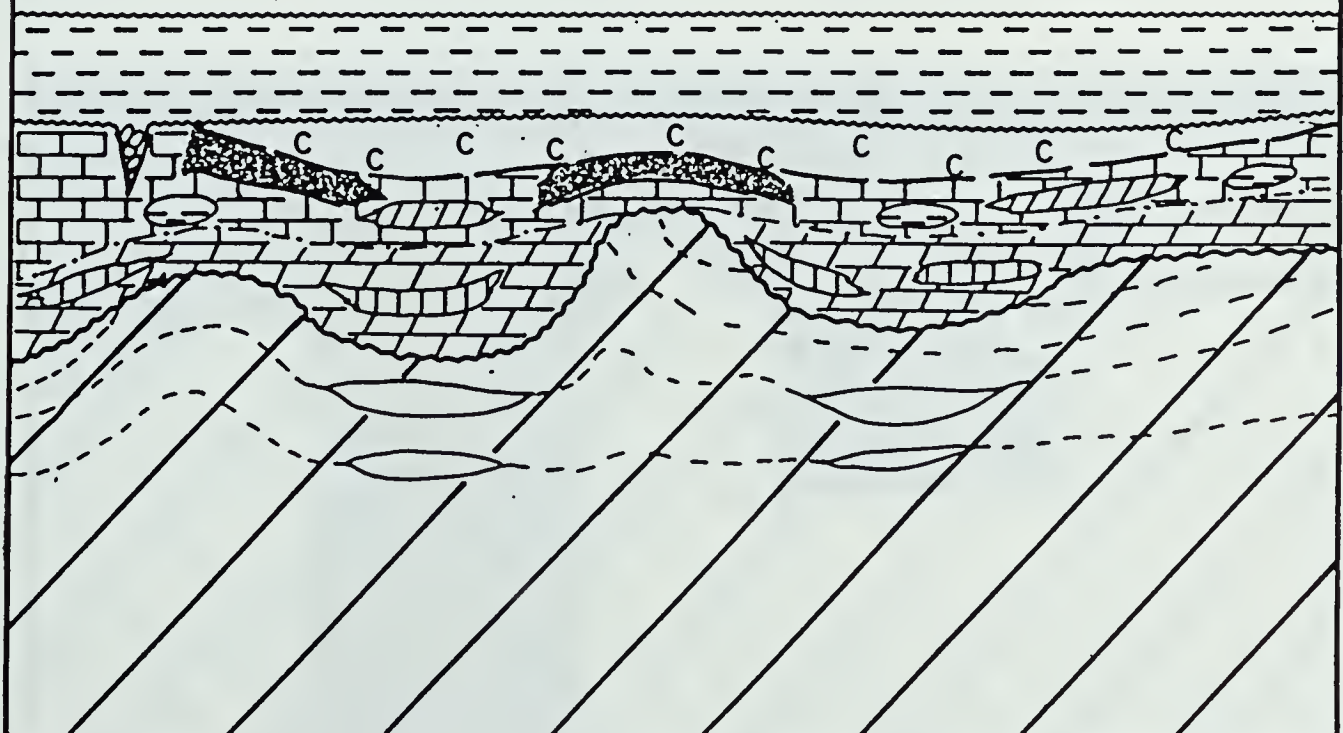
7 Development of lithofacies 2b & 2c
(recrystallization of 1st. and dolostone
replacement)

8 Development of lithofacies 7 and calcrete

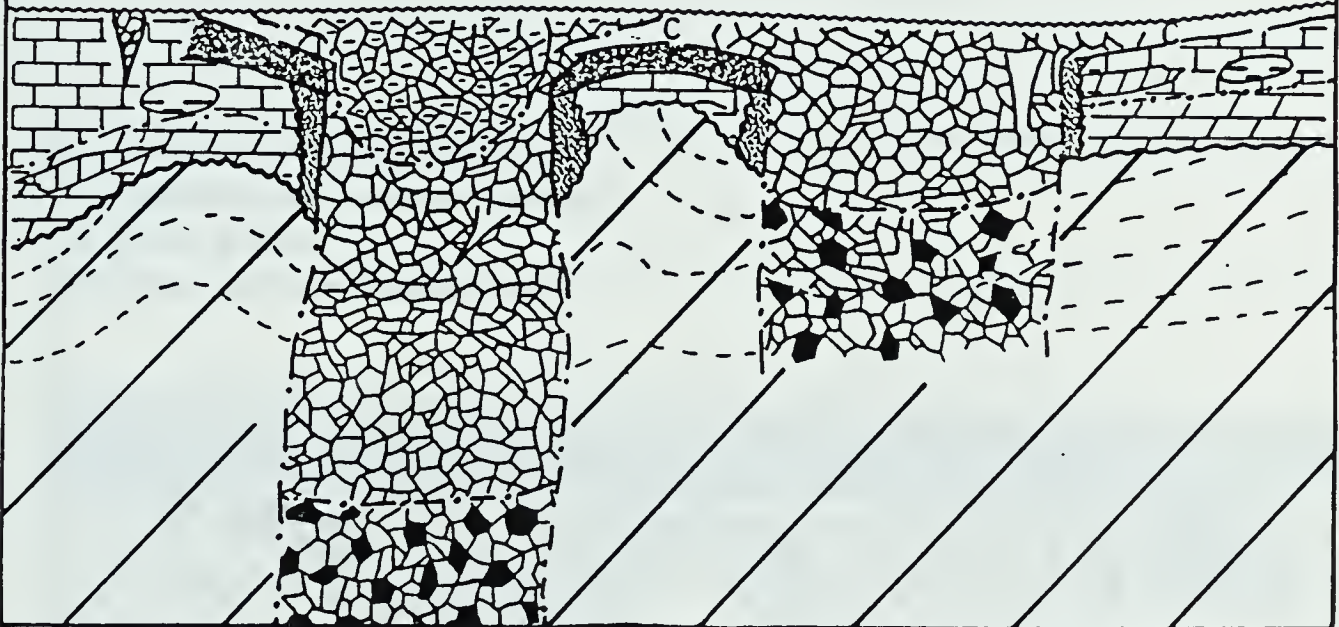


10 Peace Point Shale deposited

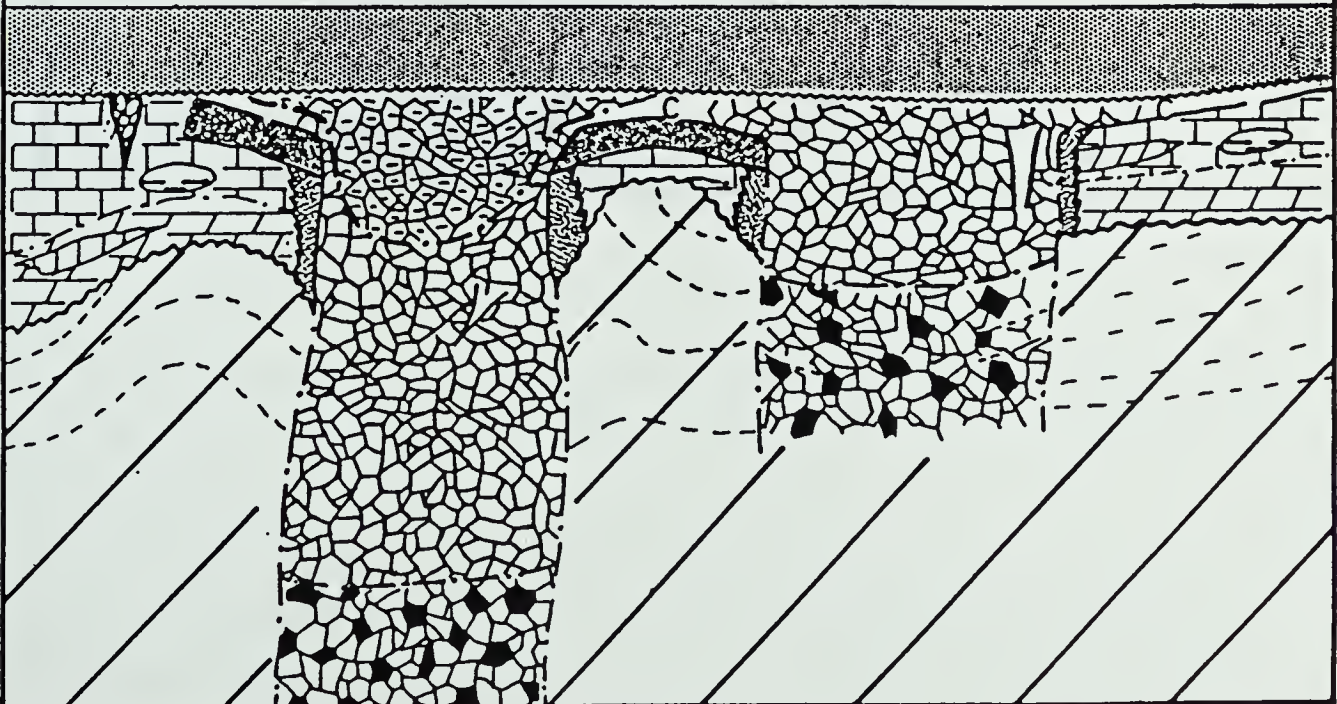
Also lines vugs of lithofacies 2b



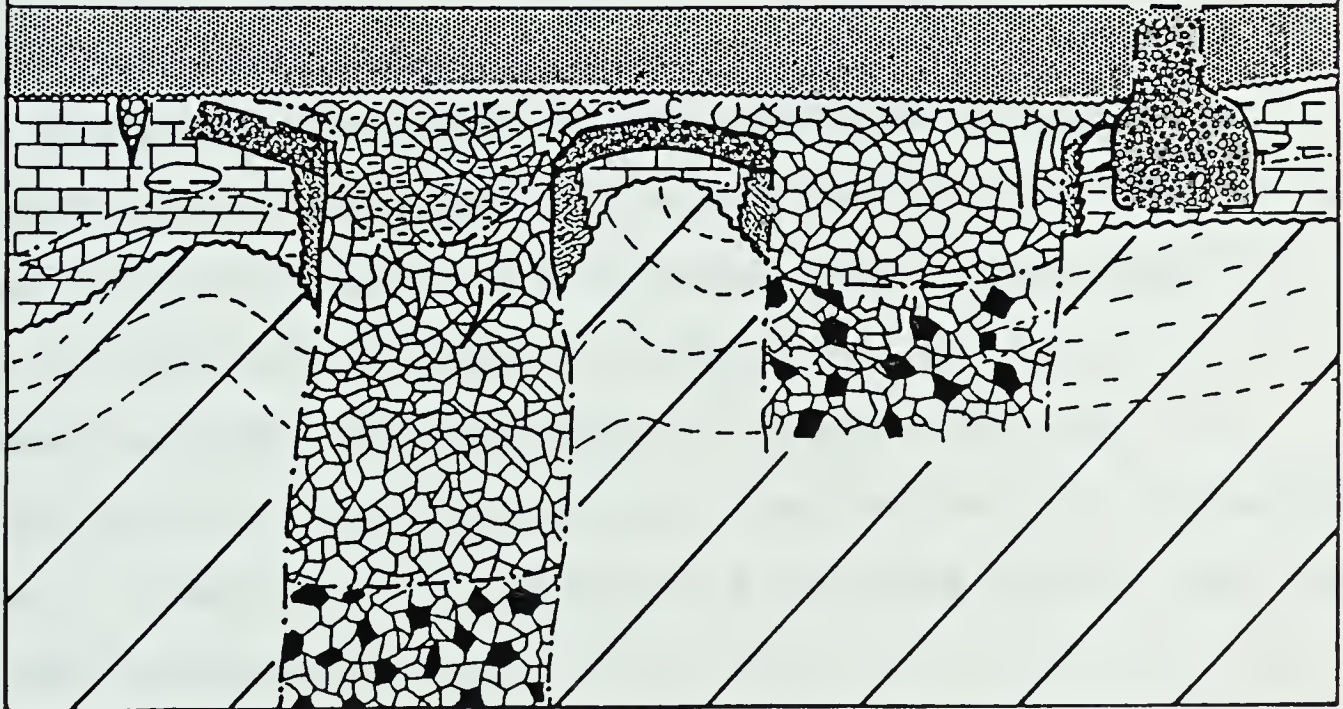
- 11 *Major collapsing event (collapse 2)*
Form lithofacies 5, 6, 8, 9.
Produce lithofacies 2d on edge of collapse
- 12 *Continued erosion at unconformity*



- 13 *Cretaceous strata deposited*
Sandstone lenses form in lithofacies 2b (not shown).



- 14 Collapse 3. Form lithofacies 10
FOLLOWING NOT SHOWN ON DIAGRAM
15 Tilting of strata
16 Erosion to present erosion surface



gradational aspect, the relative ages, relationships to unconformities, the lateral relationships between the lithofacies and the relationship to a karst/calcrete profile of the breccias, clearly suggest that lithofacies 5, 6, 8 and 9 are of a collapse origin. For reasons given, the dissolution of evaporite is thought to be the main collapsing factor. Therefore, the breccias represented by lithofacies 5, 6, 8 and 9 are best named evaporite-solution-collapse breccias.

COLLAPSE 3 - Lithofacies 2b formed prior to deposition of the Cretaceous sandstone, however dissolution of lithofacies 2b probably continued even after deposition of the sandstone. The weight of the sandstone on top of the diagenetically weakened, partially dissolved, lithofacies 2b resulted in collapse and the formation of lithofacies 10. Small, isolated collapses in lithofacies 2b, on top of bedded outcrops of lithofacies 1, demonstrate that dissolution of evaporite was not involved in this collapsing event. Lithofacies 10 represents a collapse breccia that had formed in paleo-dolines.

B. HISTORY - PEACE POINT BRECCIATION

The history of the Peace Point areas is as follows (Fig. 36):

1. Lithofacies 1 was deposited.
2. Lithofacies 1 was deformed, with anticlines and synclines produced.

3. An angular unconformity occurred on top of lithofacies 1.
4. The Slave Point Formation was deposited parallel to the unconformity.
5. Subaerial exposure of the Slave Point Formation produced extensive diagenesis, with lithofacies 2a and 3 representing the original lithology. The following features were produced as a result of subaerial exposure.
 - a. Collapse 1 - This collapse produced paleo-sinkholes in the Slave Point Formation. The breccia that formed in the paleo-dolines is lithofacies 4.
 - b. A karst profile formed, in which lithofacies 2b and 2c were produced. These two lithofacies may represent diagenesis in the upper vadose and lower vadose/phreatic environments.
 - c. A terra rossa developed at the top of the karst profile.
 - d. A calcrete horizon caps the karst profile. It may reflect climatic changes that occurred after the formation of the karst.
 - e. Meteoric water caused extensive dissolution and cave development beneath the lows in lithofacies 1, in the lenticular, upper phreatic zone.
6. The shale of the Peace Point Member was deposited. This shale fills in vugs produced in lithofacies 2b.
7. Collapse 2 - Caves in lithofacies 1 collapsed producing

collapse 2. This is the major collapsing/brecciating event in the history of the area and is responsible for many of the lithofacies.

a. Lithofacies 5, 6, 8 and 9 represent evaporite-solution-collapse breccias. Lithofacies 5 formed brecciated pockets of Peace Point Shale in lithofacies 6, 8 and 9. Lithofacies 6, 8 and 9 formed as one single collapse where there is a gradational aspect and a relative age hierarchy to the breccias. The shale of the Peace Point Member was incorporated into these breccias, increasing in abundance downward. The degree and extent of collapse was varied with some collapses being more extensive than others.

b. Lithofacies 2d formed on the edges of the collapses with a ropey and convoluted texture developed.

8. A major unconformity exists until the Early Cretaceous when calcareous sandstone (McMurray Formation) of Upper Barremian age was washed into vugs in lithofacies 2b, below the exposure surface.
9. Deposition of the Cretaceous sandstone and shale of the McMurray Formation blankets the area with the rest of the McMurray Formation being deposited from Upper Barremian to Aptian time. Shales and carbonaceous siltstones of unknown lithostratigraphic identity were deposited from Santonian to Maestrichtian time.
10. Collapse 3 - Collapse breccias, in the form of

paleo-dolines, formed lithofacies 10 during a post-Maestrichtian pre-Tertiary collapsing episode.

11. Tilting of strata and erosion to present level.

IX. CONCLUSION

The Devonian strata of Wood Buffalo National Park represents deposition in the Lower and Middle Devonian Elk Point Group and Middle to Upper Devonian Beaverhill Lake Group. The Keg River Formation was previously divided into three members in Wood Buffalo National Park; however, the presence of a regional marker (the Keg River marker) divides the Keg River Formation into two previously designated regional members, the Lower Keg River Member and the Upper Keg River Member. Therefore, in Wood Buffalo National Park the Lower Keg River Member is herein divided into units A through C, whereas the Upper Keg River Member is represented by unit D. Strata exposed at Salt River occur in the Lower Keg River Member because it lies stratigraphically below the Keg River marker, with the Keg River marker occurring in collapse breccias along the Salt River.

The Salt River strata mark an overall transgressive event that was dominated by sedimentation in a subtidal environment. The abundant and diverse fauna suggests that reefoid conditions had formed, but due to the high energy and extremely argillaceous conditions represented by both the fauna and rock lithologies the environment was not conducive to good reef forming conditions. Instead, sedimentation occurred on a high energy, bioclastic carbonate platform that marks the ancient, Devonian coastline of the La Crete Basin. A regressive event at the end of deposition of unit B is marked by a diastem produced

by subaerial exposure. Extensive diagenesis, including three phases of dolomitization and subsequent dedolomitization, is documented below the diastem, but not above it. This diastem may be the same break which produced the breccias observed at Hay Camp.

A geological history for the Salt River strata was formulated, based on the relationship of diagenesis and brecciation to the subaerial exposure surface.

The genesis of breccias at Salt River was a two-fold process. The first brecciation event formed in association to the diastem (subaerial exposure) produced at the end of unit B. Pseudoanticlines, formed in relation to a fresh water lens that existed beneath the Salt River sediments, are responsible for the first phase of brecciation. The fresh water lens caused doming of a pre-cemented sediment of the Lower Keg River Member and subsequent deformational phases, thereby producing pseudoanticlines that are represented by submarine pseudoanticlines in unit A and tepees in units B and C. These structures place significant paleoenvironmental control on the Salt River strata by showing that the Salt River strata represent a thin, probably laterally extensive belt that formed the coastline of the Devonian La Crete Basin. Intertidal to supratidal environments, represented by unit B, beneath the diastem, represent the actual coastline of the La Crete Basin produced during the Middle Devonian.

The second phase of brecciation occurred shortly after deposition of unit D, of the Upper Keg River Member, for both unit D and the Keg River marker represent the youngest strata that occur in collapse breccias along the Salt River. Evidence suggests that the collapse was associated with an unconformity that existed above unit D in the Upper Keg River Member. The previously formed pseudoanticlines acted as sites of water movement that further dissolved the Keg River strata, resulting in the collapse breccias seen. Evidence suggests that the dissolving waters were meteoric waters that were percolating down from an unconformity that existed above the Upper Keg River Member.

Evident from the Salt River and Peace River areas is the fact that multiphase brecciation at Peace Point is significantly younger and of a different nature and origin than the Salt River breccias. However, as with the Salt River area, unconformities and the relationship of diagenesis and brecciation to the unconformities provided a control for the formulation of a geologic history.

There are three major unconformities in the Peace Point study area. Previously, the Fort Vermilion Formation/Slave Point Formation contact has been described as both conformable and unconformable. This study clearly depicts the contact as an angular unconformity, where as much as 17 m of gypsum of the Fort Vermilion Formation are missing. As previously reported the Slave Point Formation/Waterways Formation contact is a subaerial exposure surface that marks

an unconformity produced between the Middle and Upper Devonian, respectively. However, this study shows that extensive diagenesis beneath this unconformity produced significant alteration of the Slave Point Formation with a karst/calcrete profile having been superimposed onto the evaporite and carbonate lithologies prior to shale deposition of the Peace Point Member. The third unconformity occurs between the shale of the Peace Point Member and the basal calcareous sandstone of Barremian to Aptian age that filled in vugs in the previously dissolved Slave Point Formation. Further dissolution of the Fort Vermilion Formation during this interval, beneath a subaerially exposed surface, produced the majority of the breccias seen at Peace Point.

Three main episodes of collapse have been identified at Peace Point. Collapse 1 occurred prior to deposition of shale of the Peace Point Member, in the early stages of development of the karst profile, with solution breccias having formed in paleo-sinkholes. Collapse 2, which resulted in most of the breccias seen at Peace Point, occurred after deposition of the Peace Point Shale, after the formation of the karst/calcrete profile, but prior to deposition of the Cretaceous sandstone. This collapse occurred as a result of dissolution of the Fort Vermilion Formation beneath lows developed between evaporite domes of the Fort Vermilion Formation; thereby forming evaporite-solution-collapse breccias. Extensive diagenesis and the lack of any

lithologies younger than the Upper Devonian Peace Point Member or older than the Barremian age sandstone and shale of the McMurray Formation suggests that the area at Peace Point was subaerially exposed from deposition of the former to the latter. Therefore, the conclusion is herein made that the Wood Buffalo National Park area was probably high and dry from the Upper Devonian to the Barremian of Early Cretaceous time. Like the Peace River Arch, undetected flanking reefs, carbonate banks and other exploration targets may fringe this northeastern high to the west. Collapse 3 occurred after the deposition of carbonaceous siltstones of Maestrichtian age, which occurred after deposition of the Upper Barremian to Aptian shales, siltstones and sandstones of the McMurray Formation. The lithostratigraphic identity of the Santonian to Maestrichtian rocks is not known. Palynomorphs, in the shale, siltstones and sandstones from collapse breccias in paleo-dolines, are indicative of the Devonian-Cretaceous contact, the McMurray Formation and the upper deposits of the Late Cretaceous.

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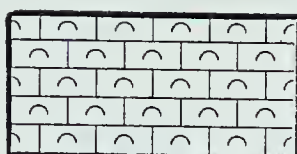
XI APPENDICES

APPENDIX 1: STRATIGRAPHIC X-SECTION
OF THE LOWER KEG RIVER MEMBER
SALT RIVER AREA, WOOD BUFFALO
NATIONAL PARK (IN POUCH)

APPENDIX 2: MEASURED SECTIONS AT SALT
RIVER, WOOD BUFFALO NATIONAL PARK

Lithology Symbols for the Keg River Sections

311



UNIT C2 - A prominent, pale orange buff to medium grey, thinly bedded to medium bedded, aphanitic to very finely crystalline, highly fossiliferous rubbly limestone.



BIOMICRITE MARKER - A recessive, light olive to pale greenish grey, rubbly, packed biomicrite to micrite.



UNIT C1 - A semiprominent to prominent, light grey to orangy buff to pale green, very thinly bedded to medium bedded, very finely crystalline, fossiliferous crinoidal, rubbly limestone.



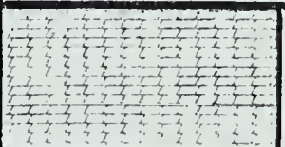
UNIT B2/B3 - Where unit B2 and unit B3 cannot be differentiated in the field.



UNIT B3 - A semirecessive to semiprominent, light to medium grey to orangy buff, thinly bedded to medium bedded, calcareous finely crystalline dolostone that contains medium to dark brown, highly argillaceous patches.



UNIT B2 - A semirecessive to semiprominent, buff to orangy buff, thinly bedded to medium bedded, calcareous very finely crystalline dolostone that contains discontinuous wispy laminations, horizontal stylolites and white dolomite filled vugs.



UNIT B1 - A semirecessive, orangy buff to cream, very thinly bedded to thinly bedded, calcareous finely crystalline dolostone that contains brown to black, subhorizontal to horizontal, irregular argillaceous laminae. Individual convex and concave, wispy laminae give the rock a rubbly, brecciated, nodular appearance.

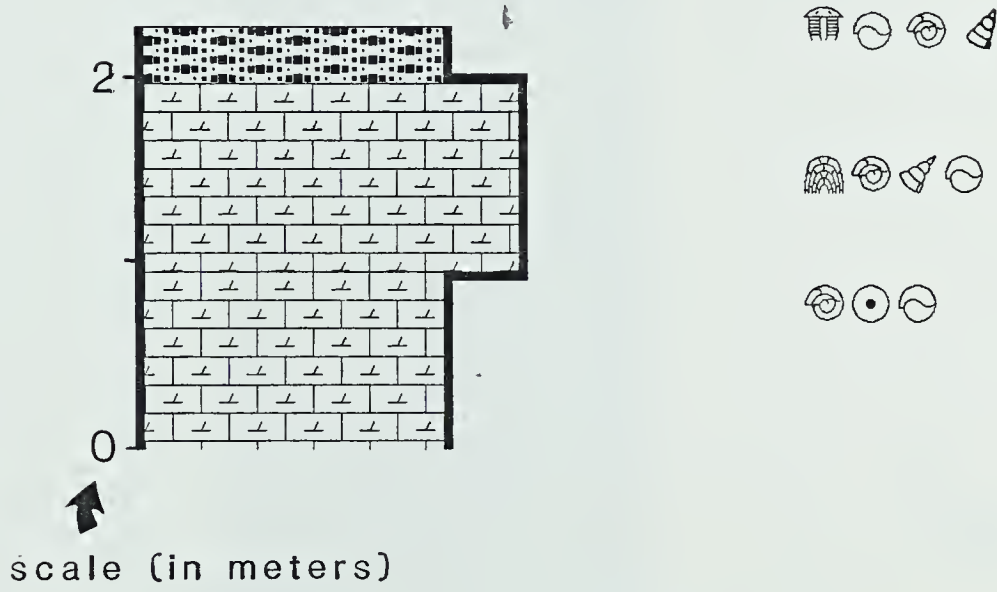


UNIT A - A recessive, medium grey to brown, thinly laminated to thinly bedded, dolomitic aphanocrystalline limestone that contains dark brown 2 wide, wispy subhorizontal to horizontal, argillaceous laminae.

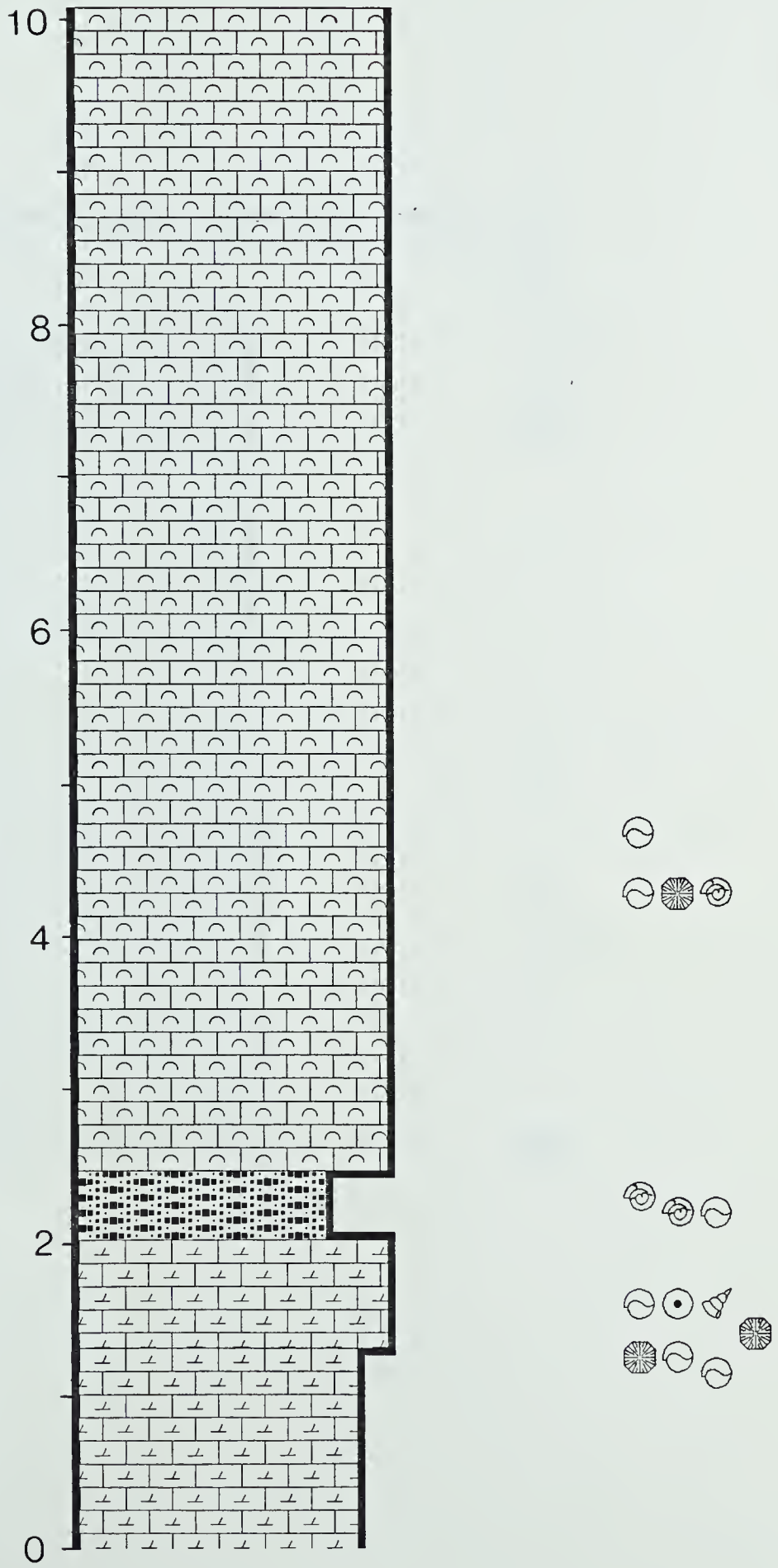
Fauna Symbols for the Keg River Sections

	brachiopods		orthocones
	bivalves		
	conical - spiral gastropods		
	planispiral - spiral gastropods		
	solitary and colonial corals		
	stromatoporoids		
	trilobites		
	crinoids		
	horizontal burrows		
	bryozoans		

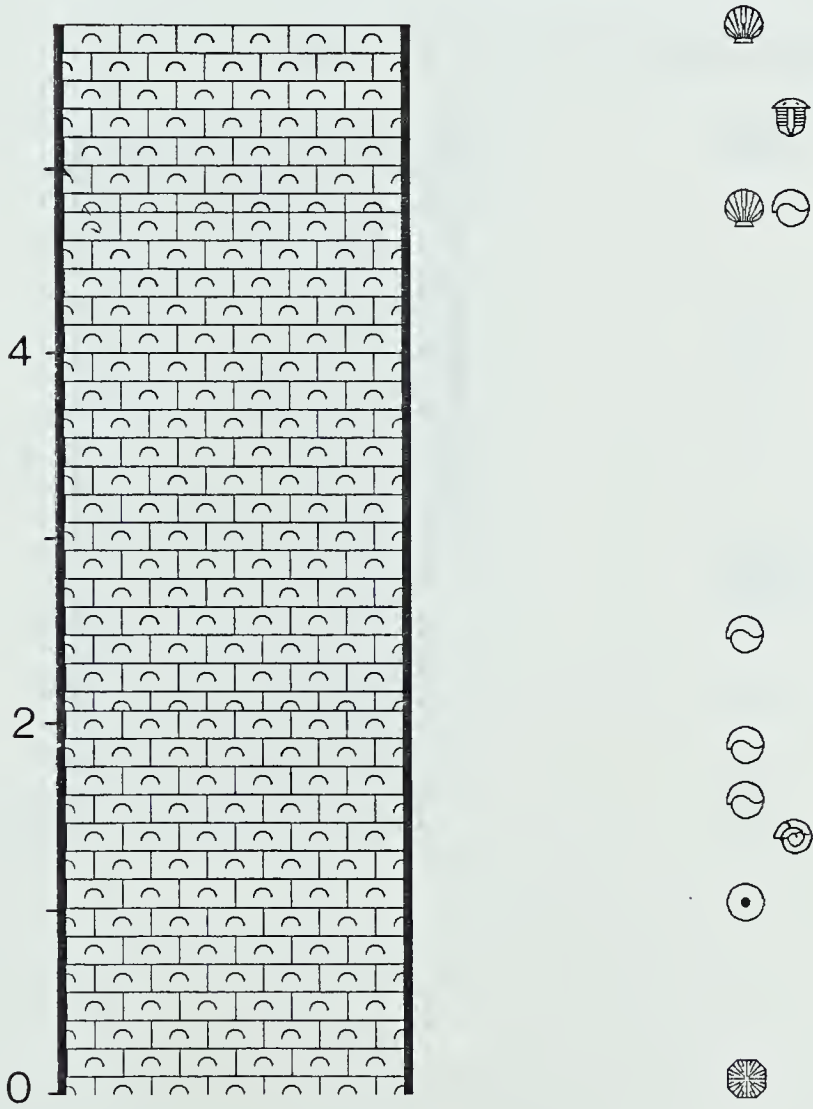
SECTION A. SALT RIVER, WOOD BUFFALO PARK



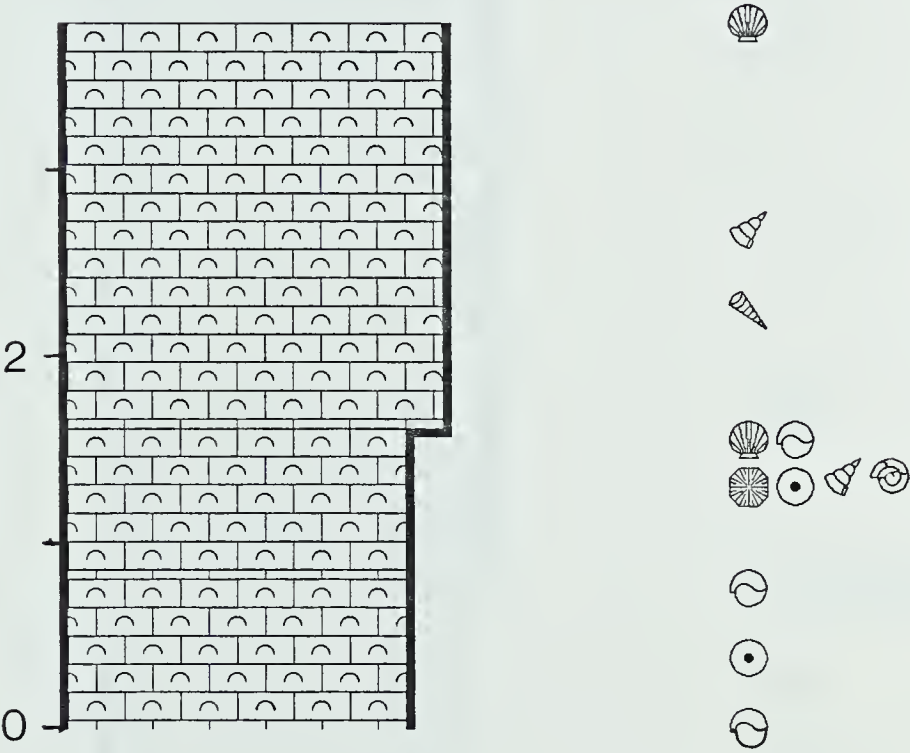
SECTION B, SALT RIVER, WOOD BUFFALO PARK



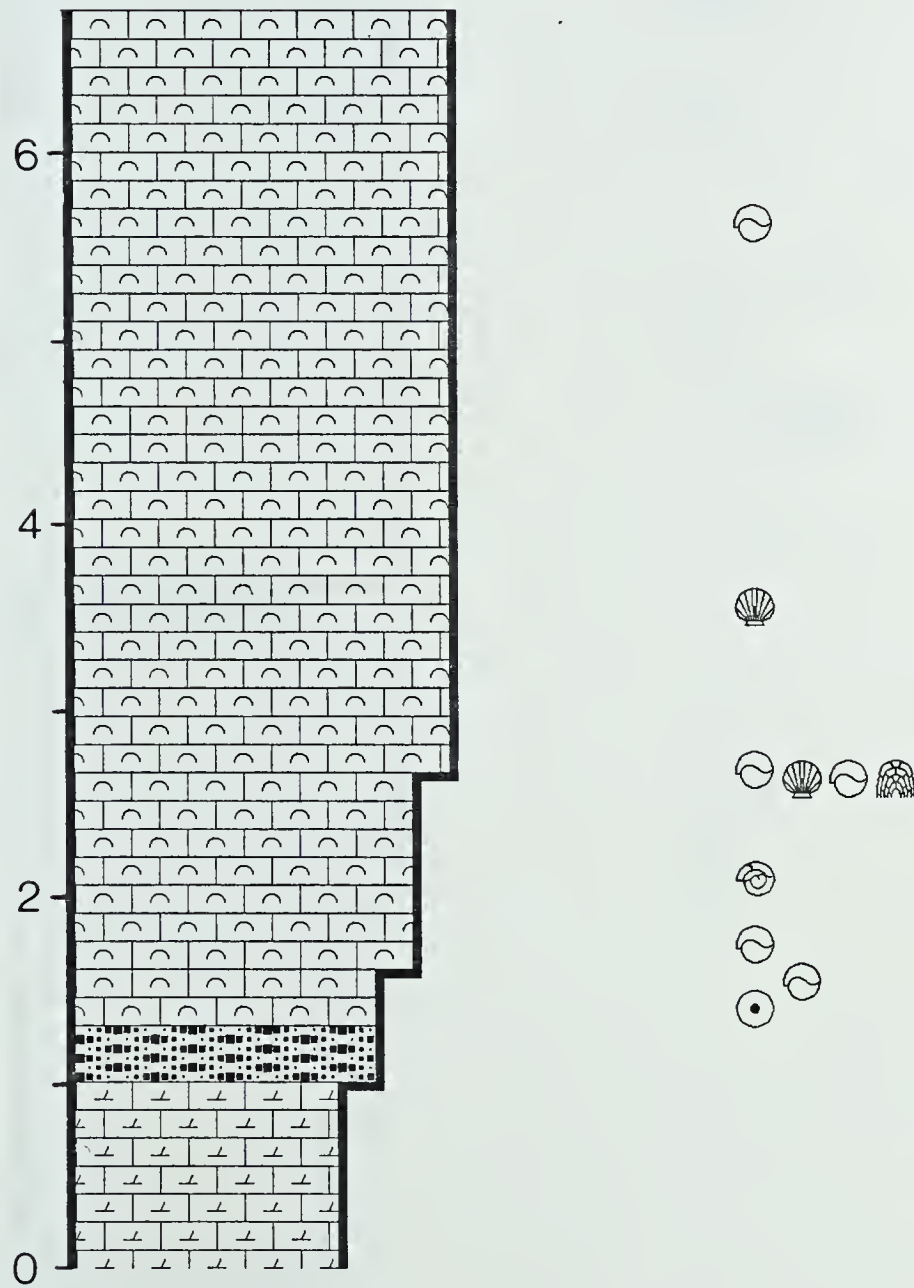
SECTION C, SALT RIVER, WOOD BUFFALO PARK



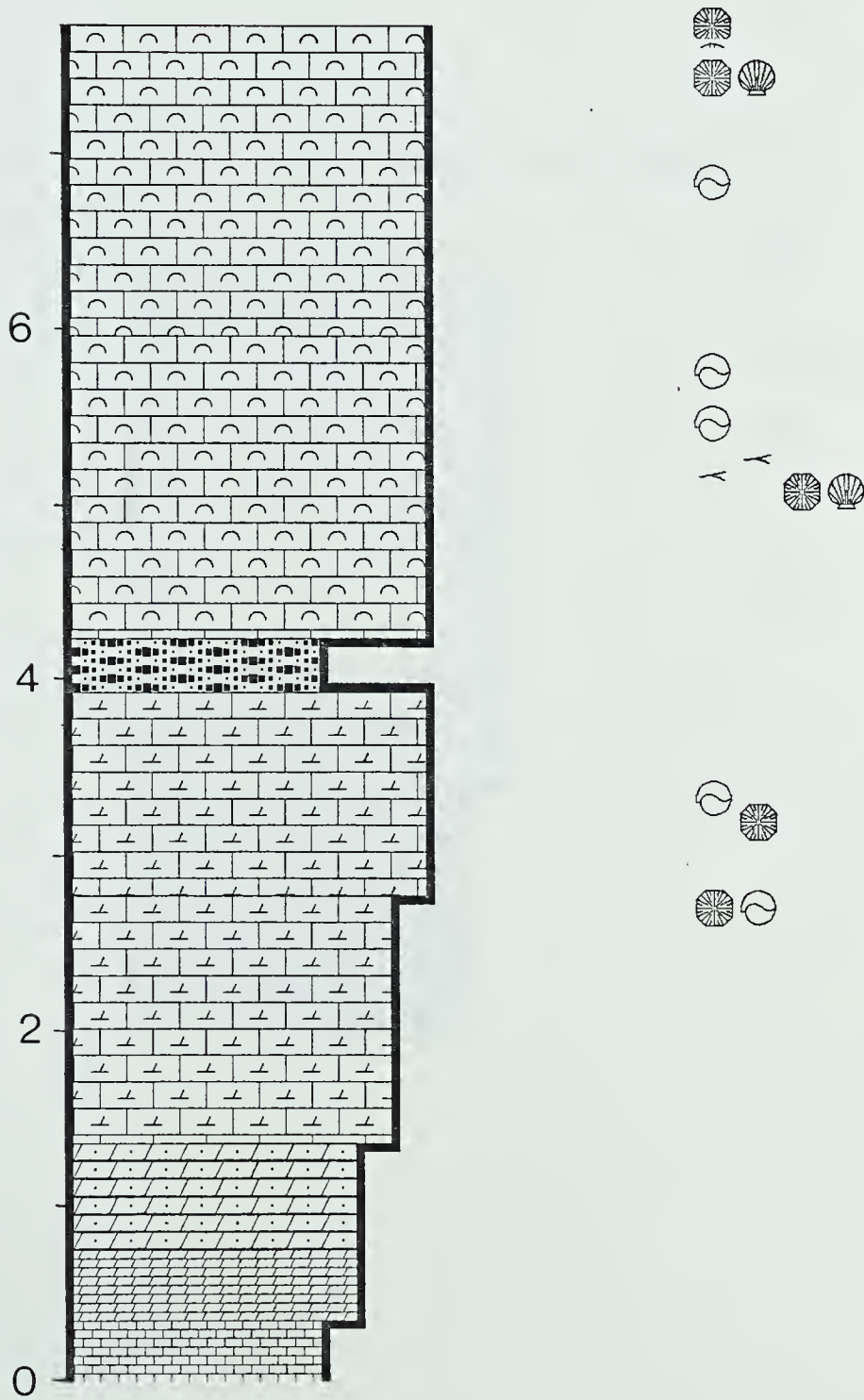
SECTION D, SALT RIVER, WOOD BUFFALO PARK



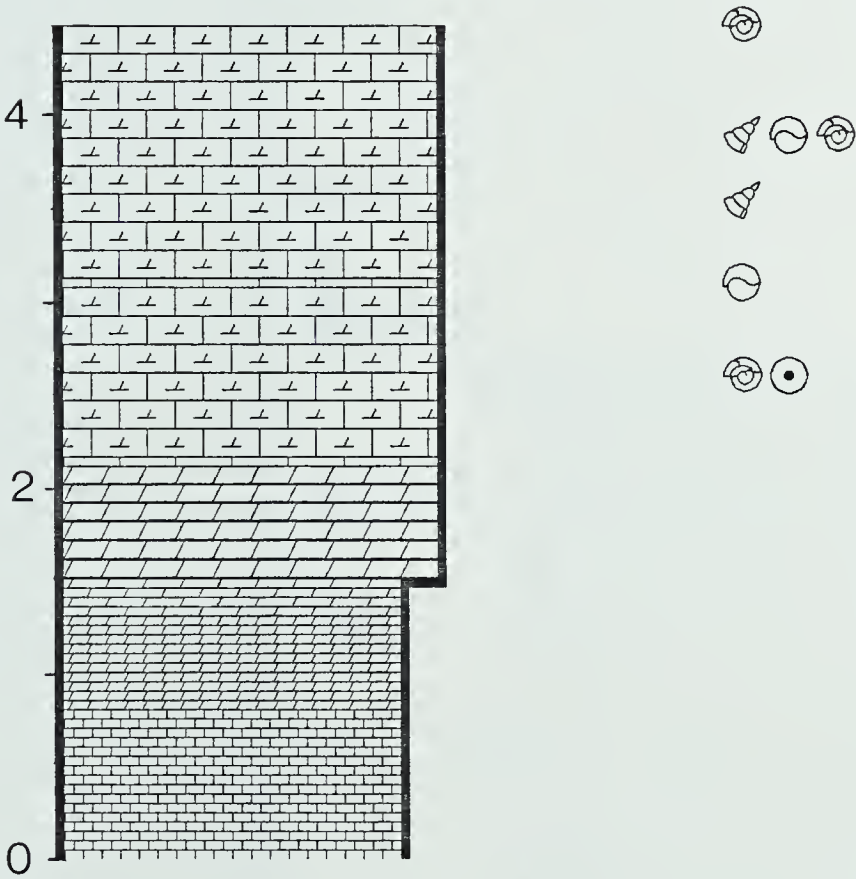
SECTION E, SALT RIVER, WOOD BUFFALO PARK



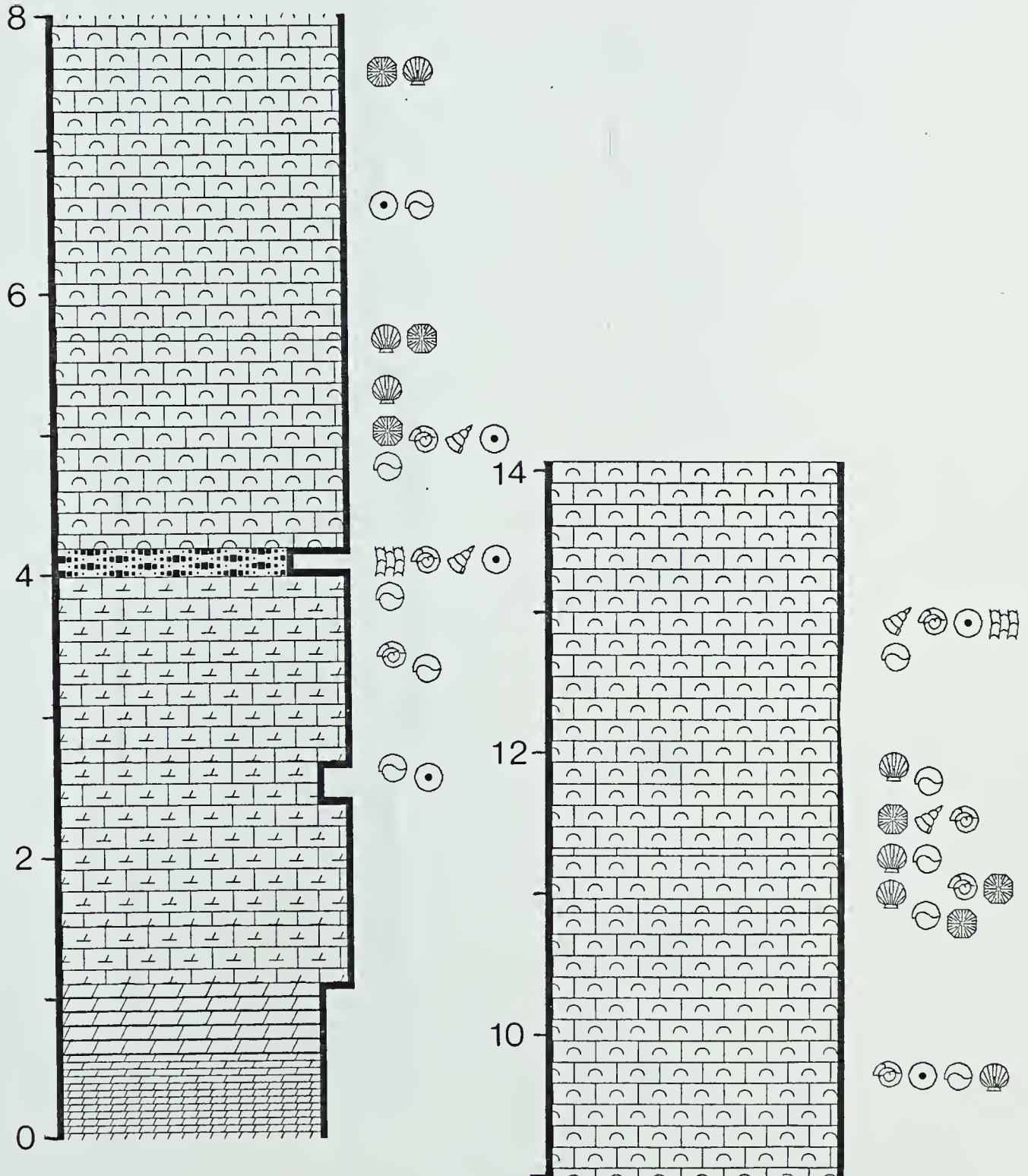
SECTION F, SALT RIVER, WOOD BUFFALO PARK



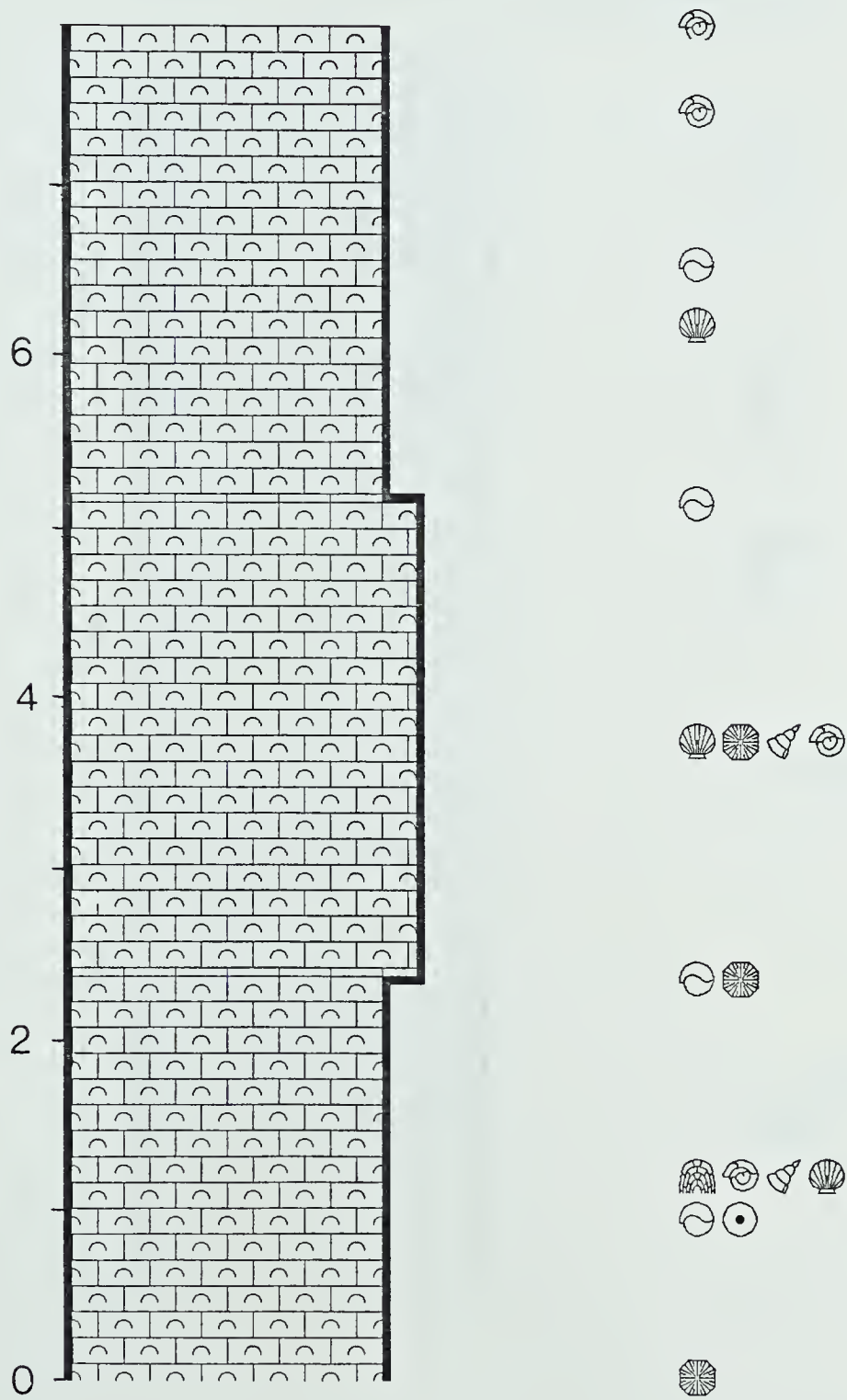
SECTION H, SALT RIVER, WOOD BUFFALO PARK



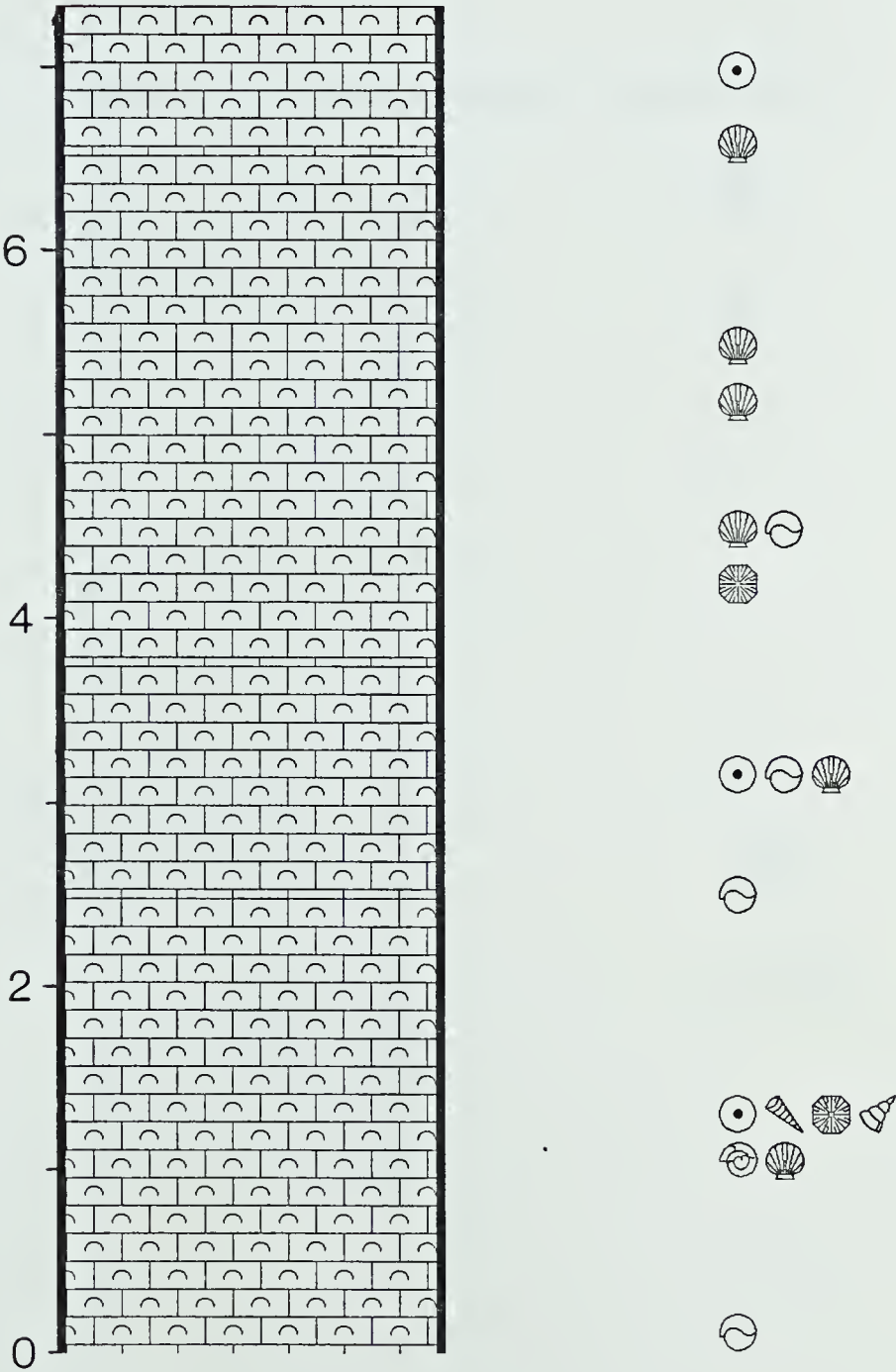
SECTION J, SALT RIVER, WOOD BUFFALO PARK



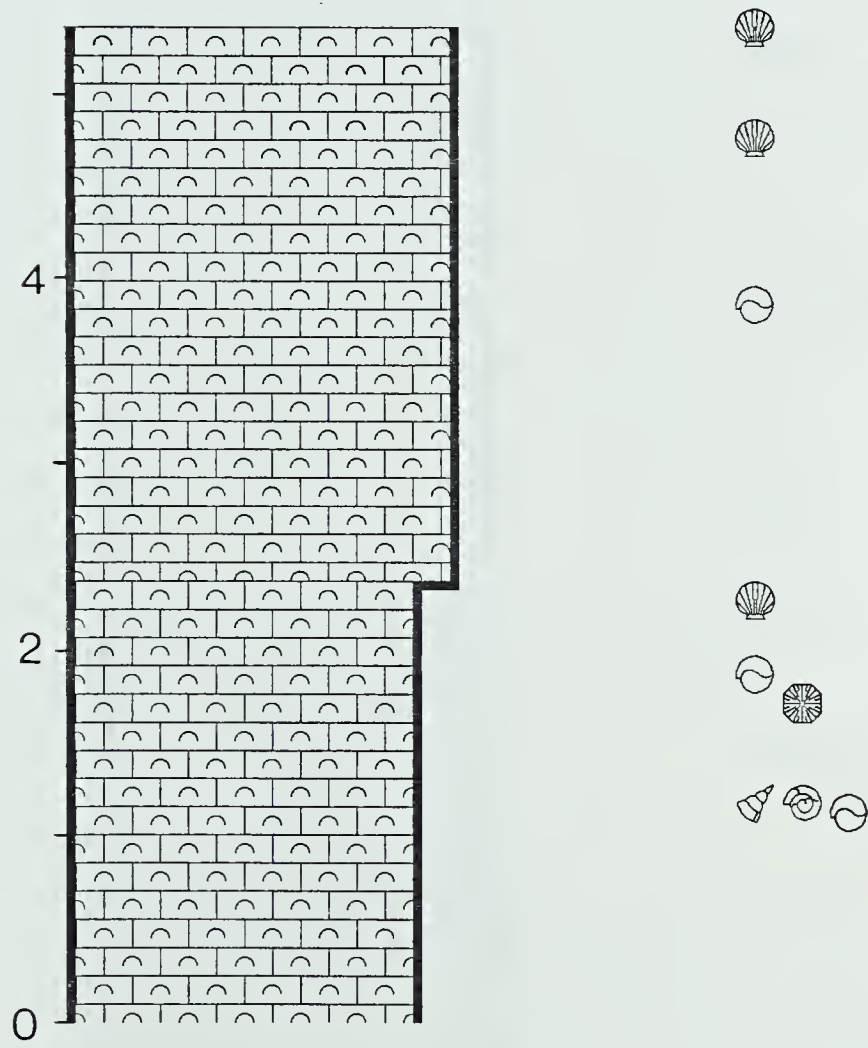
SECTION K, SALT RIVER, WOOD BUFFALO PARK



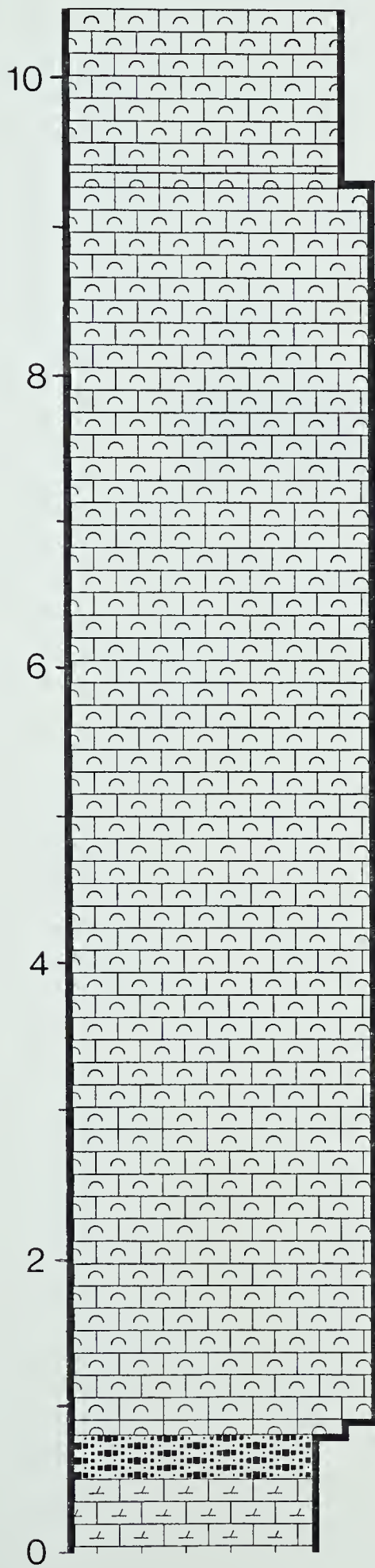
SECTION L, SALT RIVER, WOOD BUFFALO PARK



SECTION M, SALT RIVER, WOOD BUFFALO PARK



SECTION 0, SALT RIVER, WOOD BUFFALO PARK

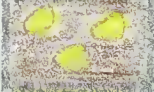


APPENDIX 3: SCHEMATIC DRAWING FROM
PHOTOGRAPHS TAKEN OF SALT RIVER EXPOSURES,
WOOD BUFFALO NATIONAL PARK

APPENDIX 4: COMPUTERIZED SCHEMATIC
DRAWINGS DIGITIZED DIRECTLY FROM
PHOTOGRAPHS TAKEN OF THE GYPSUM CLIFFS,
PEACE POINT AREA, WOOD BUFFALO NATIONAL
PARK (IN POUCH). APPROXIMATELY 600X
REDUCTION. SECTIONS RUN FROM WEST TO EAST
FROM TOP LEFT TO TOP RIGHT TO BOTTOM LEFT
TO BOTTOM RIGHT. IN SECTION 2, 0 REPRESENTS
A TALUS COVERED AREA WHERE 102M OF PHOTO
EXPAND AND DECREASE IN SCALE, THEREFORE SO
DO THE SCHEMATIC DIGITIZED DRAWINGS. TO
HELP ELUCIDATE THIS 1.5M HIGH SCALE BARS HAVE
BEEN PLACED ON THE SECTIONS.

(IN POUCH)

LEGEND FOR DIGITIZED SECTIONS

	LITHOFACIES 1
	LITHOFACIES 2a
	LITHOFACIES 2b
	LITHOFACIES 2c (massive variety)
	LITHOFACIES 2c (mixing variety)
	LITHOFACIES 2d
	LITHOFACIES 3
	LITHOFACIES 4
	LITHOFACIES 5
	LITHOFACIES 6
	LITHOFACIES 7
	LITHOFACIES 8
	LITHOFACIES 9
	LITHOFACIES 10 sandstone clasts
	LITHOFACIES 10 breccia matrix
	OVERBURDEN
	BEDDING

APPENDIX 5: PALYNOLOGICAL DATA FROM SAMPLES
FROM WOOD BUFFALO NATIONAL PARK

ALL SAMPLES PREPARED AND IDENTIFIED BY
C. SINGH OF THE ALBERTA GEOLOGICAL SURVEY,
ALBERTA RESEARCH COUNCIL

Appendix 5

Sample ZE15

Rock type: Calcareous Sandstone

Locality: Most western occurrence of sandstone. Extreme eastern end of section PE. Taken from top of calcrete horizon (lithofacies 8). From a laterally restricted collapse.

Preservation: good, mostly coal, recovery poor

Cicatricosisporites pseudotripartitus

Osmundacidites wellmanii

Alisporites grandis

Gleicheniidites senonicus

Lycopodiumsporites sp.

"*Muderongia* sp. B" Reworked? (Littoral, brackish)

Dictyophyllidites sp.

Concavissimisporites variverrucatus

Perinopollenites elatoides

Cicatricosisporites australiensis

Age: McMurray Formation U. Barremian - Aptian

Sample PEF2

Locality: From area between sections PE and PF. Originally thought area was a collapse. Appears to be collapse formed after deposition of Maestrichtian strata. Very poor exposure.

Preservation: good, mostly eval., recovery poor

Cicatricosisporites pseudostripartitus

Concavissimisporites sp. (small, granular)

Gleicheniidites senonicus

Baculatisporites comaumensis

Alisporites grandis

Concavissimisporites punctatus

Lycopodiumsporites sp.

Aquilapollenites sp.

Tricolpate (dicotyledonous angiosperm).

Age: Middle Albian. McMurray Forms mixed with *Aquilapollenites* (Santonian - Maestrichtian)

Sample PG2

Rock type: Black shale matrix of lithofacies 10

Locality: From a collapse containing lithofacies 10. Taken from
Fig. 34G.

Preservation: poor, coaly

Gleichmiidites senonicus

Trilobosporites apiverrucatus

Cicatricosisporites pseudotripartitus

Alisporites grandis

Trilobosporites triorecticulosus

Schizosporis reticulatus

Stereisporites antiquasporites

Cicatricosisporites australiensis

Concavissimisporites variverrucatus

Lycopodiumsporites reticulumsporities

Cicatricosisporites sp. damaged

Age: McMurray Formation (Upper Barremian-Aptian)

Sample PG2a

Rock type: Carbonaceous very fine grained sandstone/siltstone

Locality: From a collapse containing lithofacies 10. Sample same as Figure 35B. Clast taken from black shale matrix of PG2 (Figure 34G).

Preservation: moderate, coaly

Trilobosporites triorecticulosus

(part) *Appendicisporites bifurcatus*

Gleicheniidites senonicus

Cicatricosisporites hallei

Aequitriradites variabilis

Aequitriradites spinulosus

Age: McMurray Formation (Upper Barremian-Aptian)

Sample PG4

Rock type: Calcareous sandstone

Locality: From a collapse zone (PG4). Sample of large block of sandstone (4m high, 1.15m wide)

Preservation: moderate, very coaly

Lycopodiumsporites reticulumsporites

Cerebropollenites mesozoicus

Hymenozonotriletes mesozoicus

Gleicheniidites senonicus

Cicatricosisporites australiensis

Cicatriosisporites hallei

Age: McMurray Formation (Upper Barremian-Aptian)

Sample PG4a

Rock type: Calcareous sandstone

Locality: Of same block sample PG4 taken, only from opposite end.

Costatopertorosporites Fistulosus

Cicatricosisporites australiensis

Gleicheniidites senonicus

Cicatricosisporites halleri

Trilobosporites apiverrucatus

Foldedspore. Identi?

Age: McMurray Formation (Upper Barremian-Aptian)

Sample PG6B

Rock type: Carbonaceous siltstone

Locality: Sample taken from Lithofacies 10 collapse from a large block of carbonaceous siltstone. Block is situated in the center of the collapse (near the top of the cliff exposure).

Age: A few McMurray forms mixed with numerous *Aquilapollenites* (Santonian-Maestrichtian).

Sample PG6Eb

Rock type: Carbonaceous siltstone/very fine grained sandstone. same as PG6B.

Locality: Sample taken 2 to 3 m to left of sample PG6B, in collapse zone.

Preservation: moderate, very coaly

Trilobosporites apiverrucatus

Gleicheniidites senonicus

Callialasporites dampieri

Age: McMurray Formation (Upper Barremian to Aptian)

Sample PG6F

Rock type: Carbonaceous siltstone/ very fine grained sandstone.

Locality: Sample taken 2 m to right of sample PG6B, in collapse zone.

Preservation: poor, coaly

Gleicheniidites senonicas

Cerebropollenites mesozoicus

? *Coptospora* sp.

Cyathidites minor

Podocarpidites multesimus

Cingulatisporites sp.

Antulsporites baculatus

Age: McMurray Formation (Upper Barremian-Aptian)

Sample PG6Fa

Rock type: Carbonaceous siltstone/ very fine grained sandstone.

Locality: Sample taken from same sandstone as PG6F.

Preservation: moderate, coaly

Densosporites devonicus

? *Devonian spore*

Retustriletus sp.

Age: Devonian spores only

Sample PG6G

Rock type: Carbonaceous fine grained sandstone.

Locality: From sandstone/carbonaceous shale matrix at base of section PG6 (collapse). Unit forms matrix of Lithofacies 10.

Preservation: moderate, very coaly

Cicatricosisporites halleri

Cicatricosisporites sp. damaged

Lycopodiumsporites reticulumsporites - corroded

Callialasporites dampieri

Aequitriradites spinulosus

Appendicisporites bilateralis

Tigrisporites scurrandus

Concavissimisporites variverrucatus

Trilobosporites apiverrucatus

Verrucosisporites rotundus

Cingutriletes clavus

Age: McMurray Formation (Upper Barremian-Aptian)

Sample PG6Ha

Rock type: Carbonaceous shale

Locality: Taken from base of cliff below white dolostone breccia pipes. Same shale unit as sample PG6G taken from, only 10 m to the east.

Preservation: moderate, coaly

Ancyrospora pulchra

Verrucosisporites sp.

Densosporites devoniccus Devonian

Geminospora sp.

Apiculatisporis microconus

? Devonian spore

Trilobosporites tribotrys Lower Cretaceous

Schinzosporis reticulatus

Cicatricosisporites sternum

Age: McMurray froms are mixed with Devonian spores. Mixed strata at the Cretaceous/Devonian contact. Lower Cretaceous forms are Barremian to Aptian in age.

Sample PG6Ma

Rock type: Carbonaceous very fine grained sandstone.

Locality: From section PG6m

Preservation: moderate, coaly

Devonian spore

Relusotriletes sp.

Denosporites devonicus

Apiculatisporis sp.

Aequitriradites sponulosus

Lower Cretaceous

Age: McMurray forms are mixed with Devonian spores. Mixed strata at the Cretaceous/Devonian contact. Lower Cretaceous forms are Barremian to Aptian in age.

Sample PG6Mc

Rock type: Carbonaceous siltstone/very fine grained sandstone

Locality: From collapse zone. Extremely poor exposure. Sample suggests recent collapse.

Preservation: moderate, coaly

Lecaniella Foveata

Acanthotriletes sp. (Devonian)

Lycopodiumsporites sp.

Micrhystridium sp. (very small) Acritarch Brackish

Pterosphaeridia sp. (very small, hyaline crests) Acritarch
brackish

Modern *Alnus*

Modern *Picca*

Retusotriletes sp. (Devonian)

Bacchidinium polypes, marine, cretaceous dinoflagellate

Restusotriletes (Devonian)

Age: Devonian, Cretaceous and modern forms.

Sample PG6Me

Rock type: Carbonaceous very fine grained sandstone

Locality: From collapse zone. Same as PG6Mc, only taken from sandstone approximately 2 m above.

Preservation: moderate, mostly coaly

Gleicheniidites senonicus

Cicatricosisporites potomacensis

Taurocusporites segmentatus

Concavissimisporites variverrucatus

Trilete spore (small)

Cicatricosisporites pseudotripartitus

Lygodiosporites sp. A

Cicatricosisporites australiensis

Trilobasporites apiverrucatus

Contignisporites glebulentus

Trilobosporites tribotrys

Aeguitrinadites spinulosus (damaged)

Cingutritetes clavus

Arcellites - *A. incipiens* - Lower Mannville only

Minerisporites

Age: McMurray Formation (Upper Barremian-Aptian)

Sample PH9

Rock type: Carbonaceous very fine grained sandstone.

Locality: Taken from shale wedge in section PH (figure 24c).

Poor exposure.

Preservation: moderate, coaly

Lycopodiumsporites sp.

? Flat circular body

Trilogosporites tribotrys - Lower Cretaceous

Verrucosisporites sp. ?

?Acritarch

Micrhystridium sp.

Lophotriletes sp. (Devonian)

Trilete spore - corroded

Dibolisporites echinaceus (Devonian)

Concavissimisporites sp. (badly corroded)

Densosporites sp. (Devonian)

? Reticulate hyaline body

Age: McMurray forms mixed with Devonian spores. Mixed strata at the Cretaceous/Devonian contact.

Sample PH9f

Rock type: Sample of black shale

Locality: From shale/sst wedge at section PH (figure 24c). Very poor exposure, recessive.

Preservation: moderate, mostly coal

Cicatricosisporites australiensis

Cicatricosisporites abacus

Cingulotriletes clavus

Gleicheniidites senonicus

Deltoid oспорa sp.

Arcellites incipiens (Lower Mannville only)

Paxillitriletes sp. (part only)

Age: McMurray Formation (Upper Barremian-Aptian)

Sample PH9g

Rock type: Calcareous sandstone from black recessive shale

Locality: From shale/sandstone wedge at section PH (Figure 24C).

Preservation: moderate, mostly coal

Stereisporites antiquasporites

Classopollis classoides

Cicatricosisporites sp.

Dark object with dense hair

Gleicheniidites sp. (small)

Trilete spore, small circular, verrucate

Gleicheniidites senonicus

Cingutriletes clavus

Cicatricosisporities australiensis

Forminispora asymmetricus

Age: McMurray Formation (Upper Barremian-Aptian)

Sample K2

Rock type: Calcareous sandstone

Locality: Sample taken from Cretaceous outcrop. Same as Figure 35A. From side of doline.

Preservation: moderate, mostly coal

Cinguliriletes clavus

spherical, small body with fine hairs

Microspore

Trilobosporites trioreticulosus

Undulatisporites sp.

Cicatricosisporites australiensis

Alisporites bilateralis

Gleichenidites senonicus

Foraminisporis asymmetricus

Cyathidites minor

Trilobosporites reticulatus

Lecaniella Foveata

Cicatricosisporites sp.

Classopollis classoides

Concavissimisporites variverrucatus

Triporoletes radiatus

Trilobosporites apiverrucatus

Lycopodiumsporites reticulumsporites

Cicatricosisporites minor

Paxillitriletes fairlightensis

Arcellites incipiens (Lower Mannville only).

Age: McMurray Formation (Upper Barremian - Aptian)

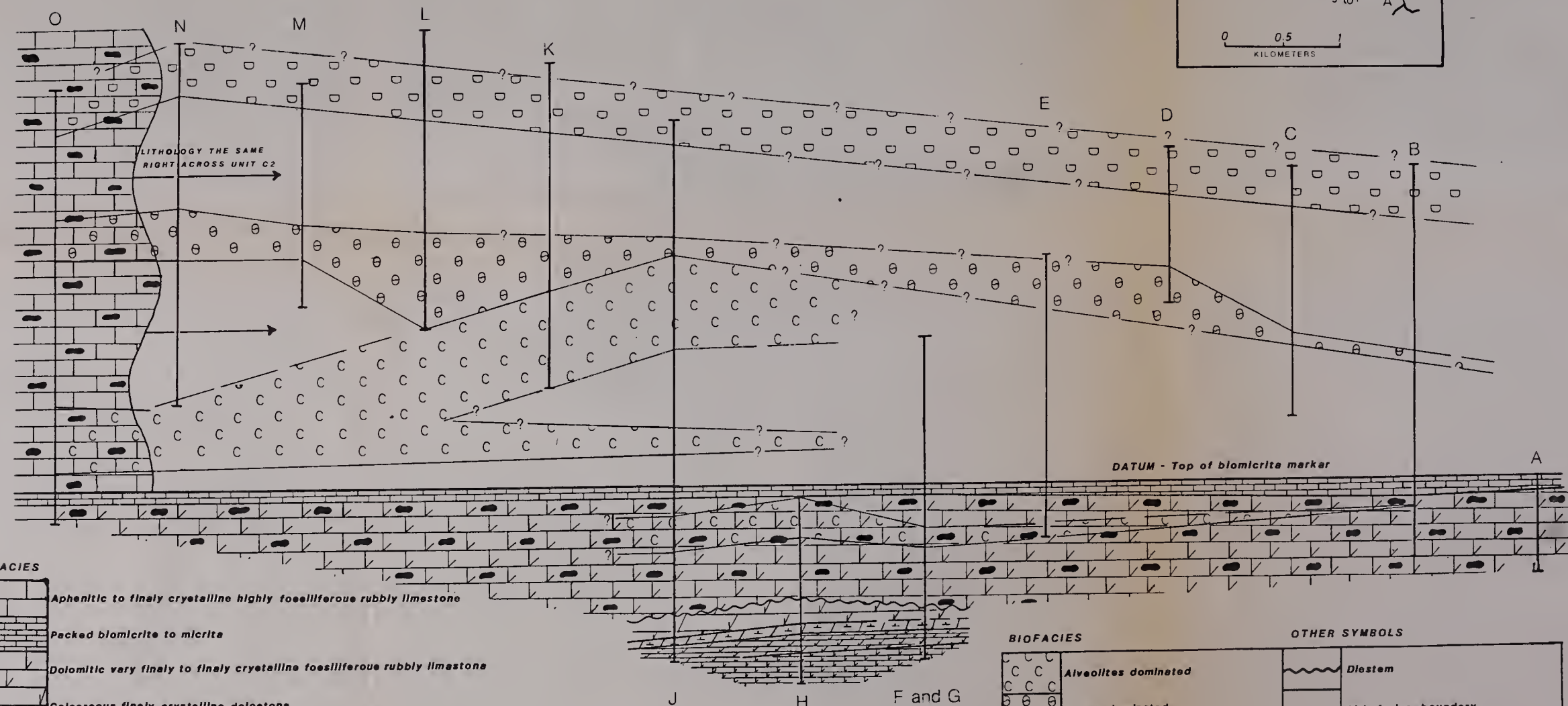
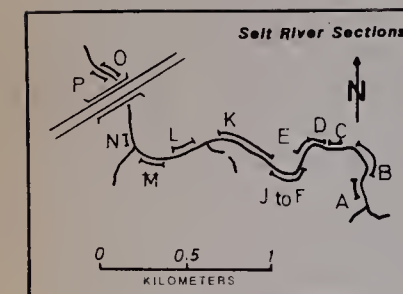
Sample K3

Rock type: Quartzitic sand non-lithified.

Locality: From same doline sample PH9g collected.

Flora: Barren

Correlation of Sections of the Keg River Formation Salt River-Wood Buffalo National Park



LITHOFACIES

	Aphenitic to finely crystalline highly fossiliferous rubbly limestone
	Packed biomicrite to micrite
	Dolomitic very finely to finely crystalline fossiliferous rubbly limestone
	Calcereous finely crystalline dolostone
	Calcereous very finely crystalline dolostone
	Calcereous aphenocrystalline dolostone
	Dolomitic aphenocrystalline limestone

BIOFACIES

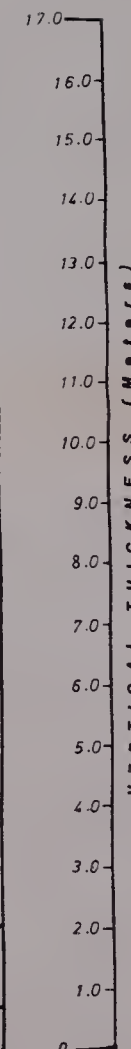
	Alveolites dominated
	Atrypa dominated
	Bivalve dominated
	No characteristic fauna

OTHER SYMBOLS

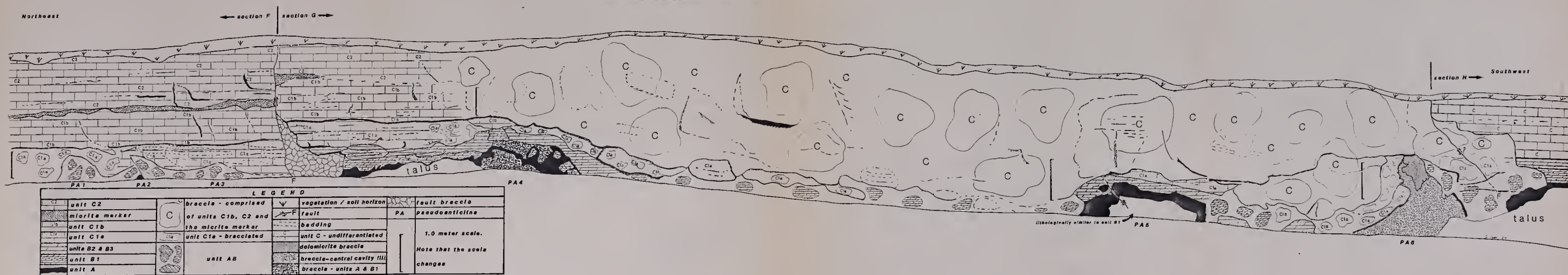
	Diastem
	Lithofacies boundary
	Biofacies boundary
	Questionable boundary

STATIGRAPHIC UNITS

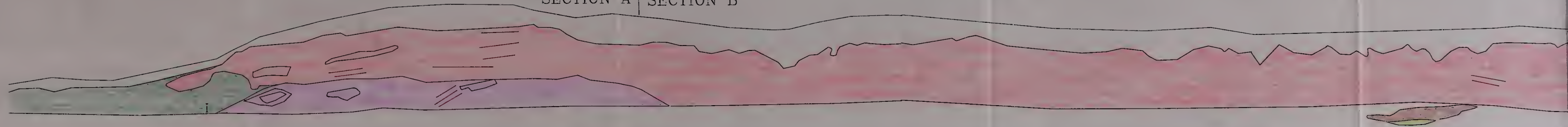
LOWER KEG RIVER MEMBER	C2	Unit C
	C1	
	B3	
	B2	
	B1	Unit B
		Unit A



LITHOSTRATIGRAPHIC AND STRUCTURAL RELATIONSHIPS OF THE LOWER KEG RIVER MEMBER
SALT RIVER WOOD BUFFALO NATIONAL PARK



SECTION A | SECTION B

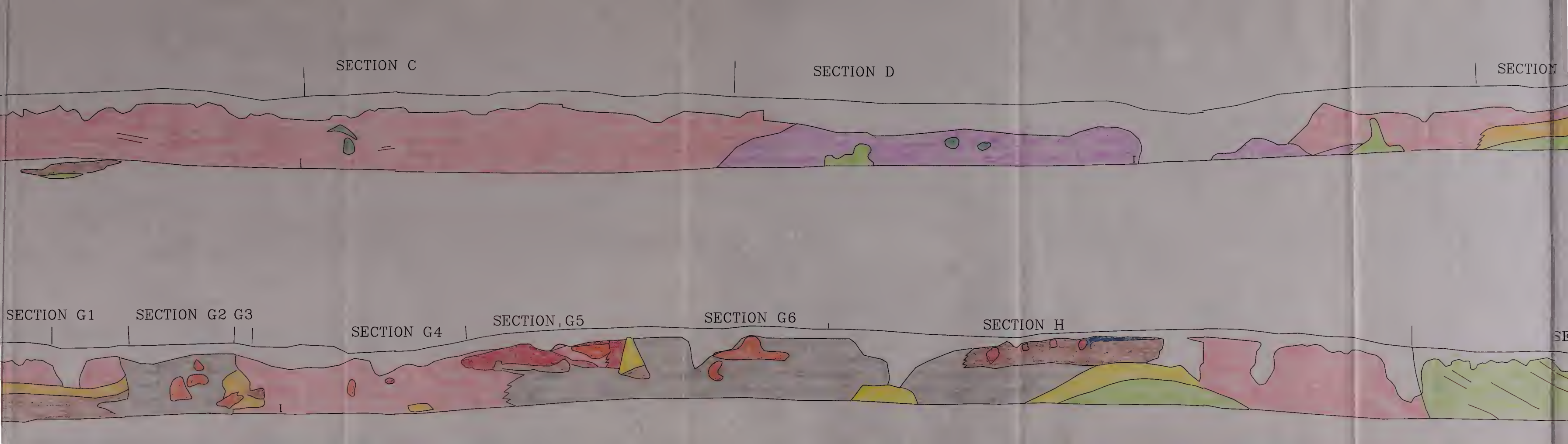


SECTION G1

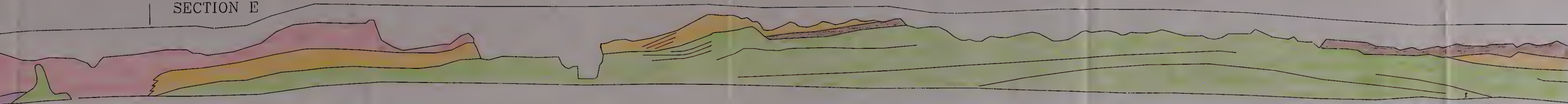
SECTION

SECTION F

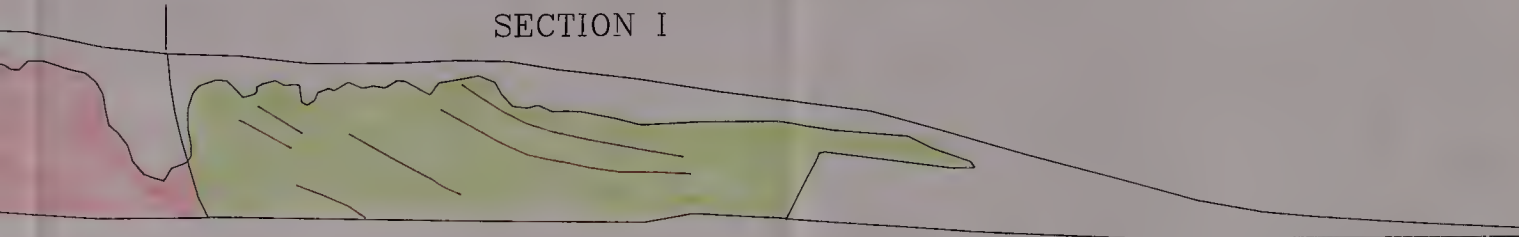




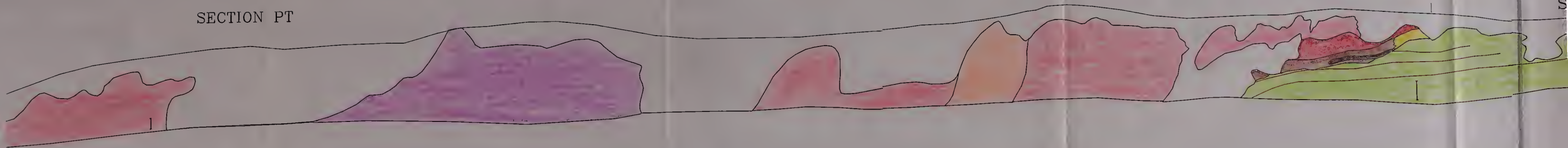
SECTION E



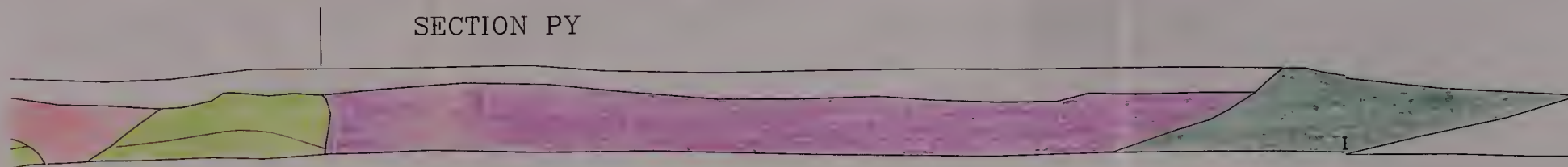
SECTION I



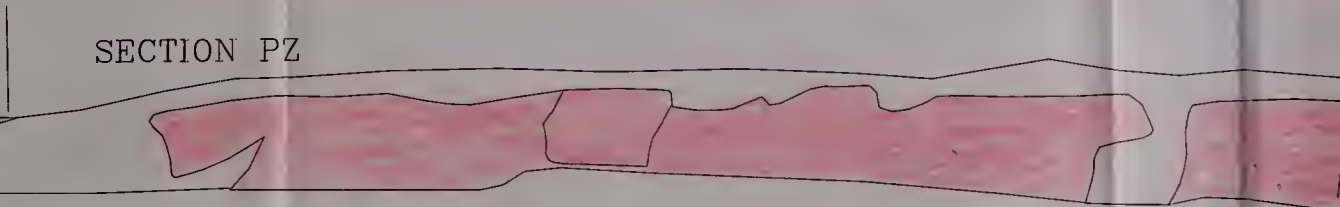
SECTION PT



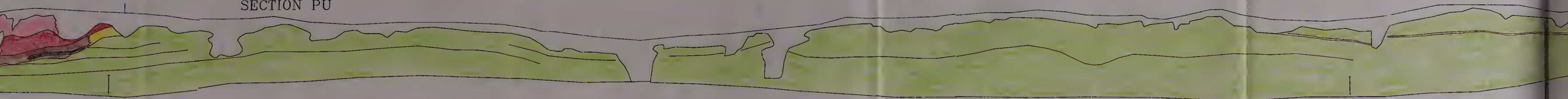
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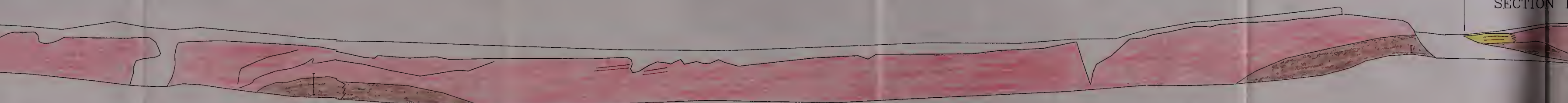
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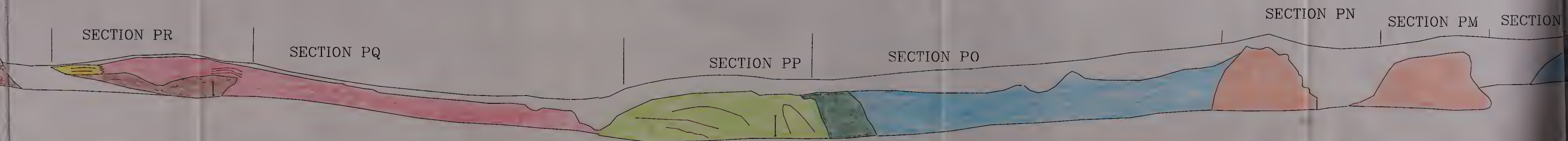
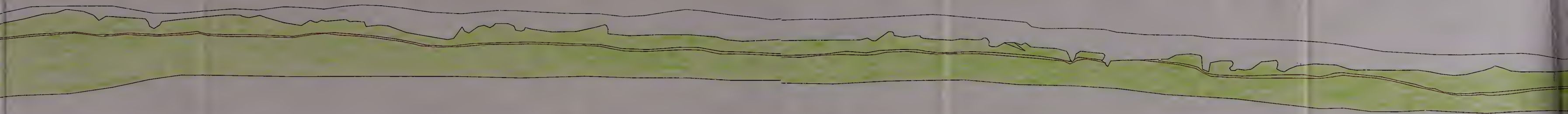


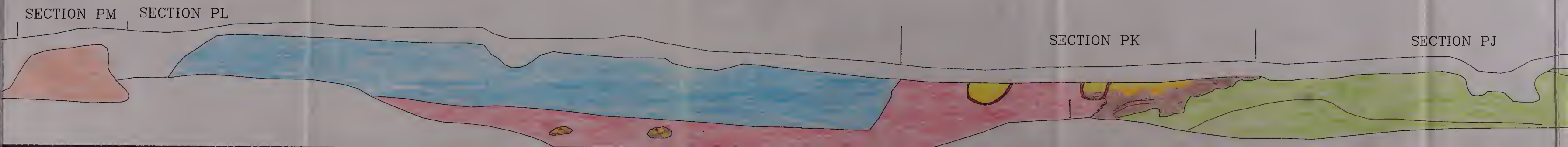
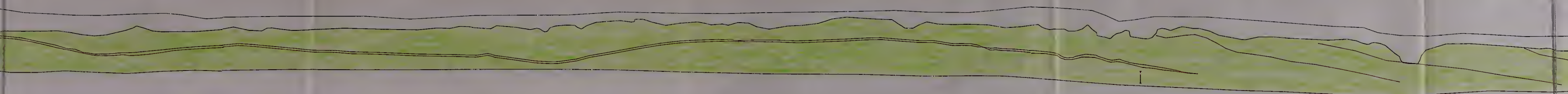
SECTION PU



SECTION I





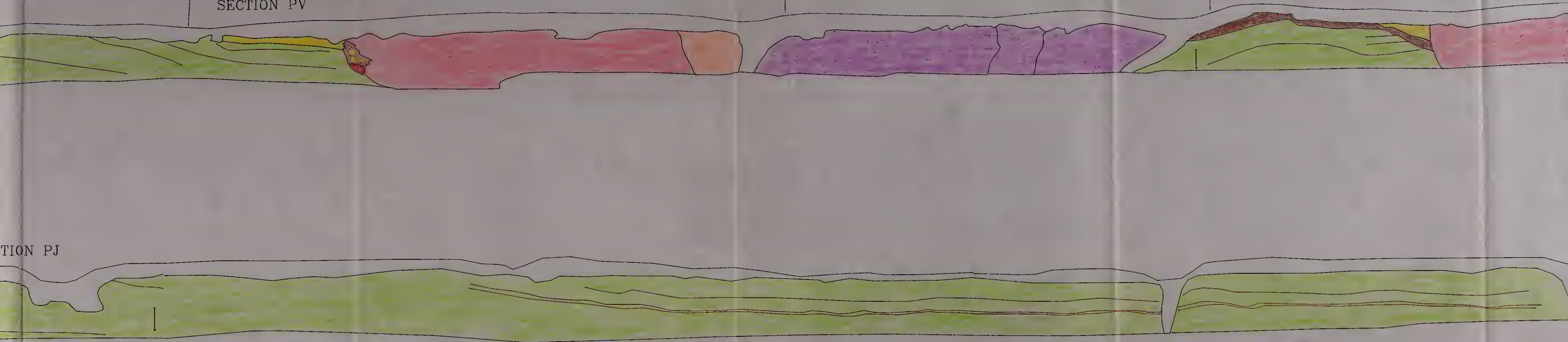


SECTION PV

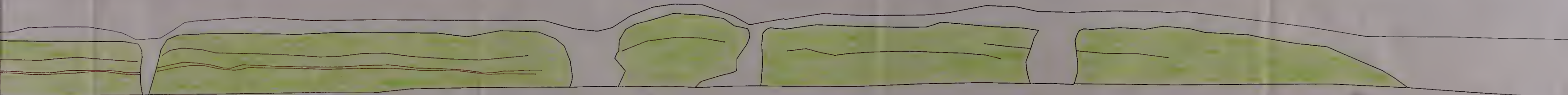
SECTION PW

SECTION PX

TION PJ



SECTION PX



B30419